Medium-term response of lowland river reaches to changes in upland land use

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In memory of Bernard Mount,
whom I am sure would have enjoyed
the 'debate'.

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ABSTRACT

The afforestation of upland areas in the UK has been the largest rural land use change this century. As a consequence of afforestation the bed load yields of upland catchments have been shown to increase substantially. The upland catchment bed load is transported downstream, through the transfer zone to the lowland reaches where it may enter storage in barforms, cause flow diversion and cause channel instability. This study investigates the Afon Trannon, a river in mid-Wales whose upland catchment was afforested between 1948 and 1978 and whose lowland channel is currently unstable. Historical rates of channel change in the lowland channel have been quantified using aerial photogrammetry in ERDAS Imagine GIS. Channel instability is shown to increase significantly between 1963 and 1976, some 15 years after upland catchment afforestation. However, upland catchment bed load yields are shown to be of low magnitude (up to an estimated maximum of 6.35 t km^{-2} yr^{-1}) and incapable of producing the high medium-term lowland channel change rates observed (up to 1.88 m yr^{-1} between 1963 and 1976). Contemporary channel DTMs constructed from field survey data have allowed the construction of a contemporary lowland channel sediment budget. Local inputs of bed load from composite bank erosion are shown to dominate in the budget and reaches of maximum instability are shown to correspond with the location of these composite banks. Additionally, flood magnitude and frequency are shown to have increased since 1988 from a maximum stage of 1.50 m between 1969 and 1988 to a maximum stage of 2.23 m between 1989 and 2000. A conceptual model is presented in which the medium-term instability of the lowland Afon Trannon is suggested to be triggered by local lowland bed aggradation as a result of elevated upland catchment bed load yields and a risk assessment diagram provides advice to river engineers and fluvial geomorphologists interested in assessing the potential stability of lowland rivers whose upland catchments have been afforested.

1. INTRODUCTION

1.1 Background

Rivers act to convey their sediment and water inputs, but are also products of that conveyance. Their form and stability can be heavily influenced by change in the inputs and the nature of their transport through the fluvial system. Consequently, fluvial geomorphologists have become increasingly interested in understanding the interaction between sediment movement, flow of water and how it affects the river channel form.

Significant change in the inputs to the fluvial system essentially originate from two sources:

1. Natural changes – e.g. change in rainfall patterns, landslides
2. Man-induced changes – e.g. water extraction, land use change

Both have the potential to cause the fluvial system to respond morphologically. However, the recent upsurge of public cognisance about the potentially damaging impact of anthropogenic activity in the landscape has meant that increasingly attention is being given to evaluating the impact of human-induced changes.

The concept that river channels adjust to meet their flow conditions is a long standing one. Indeed, the term dominant discharge has been coined as descriptive of the flow which achieves the most morphological work (Wolman and Miller, 1960). This flow is commonly taken to be the flow which just fills the available channel cross-section area (bankfull discharge). However, strong debate exists about whether this is the most effective discharge, and its recurrence interval (Knighton, 1988, pg. 164). What is clear, however, is that an increase in the magnitude and frequency of competent discharges (i.e. those flows capable of bed load transport and hence morphologic change) requires that the channel form responds as a result of bed load conveyance, and that this adjustment will be contained within in the morphologic parameters of the river – commonly the planform, cross section or a combination of both.

Morphologic change in the fluvial system results from sediment movement from areas of erosion to those of deposition, however not all sediment carried by the river is involved in the processes of morphologic change. River sediment can be subdivided into two types:

1. Suspended sediment, which is composed wash load (the fine sediment fraction whose particle settling velocities are so low that they are transported in suspension at approximately the same velocity as the flow).

2. Bed load, generally composed of material of sand size or greater, whose transport rate is a function of the transporting capacity of the flow.

It is the bed load fraction which has been shown to be the major influence on the morphology of channels (Leopold, 1992; Martin and Church, 1995) and it is therefore of particular interest when attempting to understand the impact of anthropogenic activity on river morphology. Specifically, the nature of deposition and storage of bed load within the river channel is of crucial importance. The growth of bar forms diverts flow and is able to cause significant erosion of the channel banks and lateral migration of the channel. In UK lowland rivers storage can occur at a range of scales. At the fine end of the scale exist micro and meso-scale bed forms, such as pebble clusters, which contain only small volumes of bed load and which persist for short periods and have little morphological impact on river form. At the coarse end are macro and mega-scale bar forms, such as bar forms and bar assemblages. These may contain very large volumes of bed load and persist for long periods. Increases or decreases in their size or change in their position through time can produce significant morphologic change in the channel form. Where bar growth is particularly rapid the river channel may display significant laterally instability. Such within-channel increases in storage occur when either:

1. The bed load supply to the channel is increased
2. The conveyance capacity of the river is decreased

Consequently, where anthropogenic activity has acted to significantly increase bed load inputs to the fluvial system the potential exists for lateral channel instability in regions of bed load deposition.

1.2 The fluvial system

Schumm (1977) simplified the fluvial system by dividing it into three separate zones (figure 1.1). In the upper reaches a zone of production generates sediment inputs to the system. The sediment is then transported through a transfer reach where it is deposited in a zone of deposition. Schumm recognised that the diagram is an over simplification of the fluvial system as all three processes operate within all three zones, however the dominance of the different processes in different zones of the fluvial system is well represented by the diagram.

Figure 1.1 The simplified fluvial system of Schumm (1977).

Figure 1.1 generates important inferences for the potential impact of anthropogenic activity in the upland catchments of rivers on lowland reaches. Specifically, it implies that a man-induced increase in upland catchment bed load yields would be expressed morphologically not in the upland channel reaches, but rather in the lowland channel reaches where the material enters storage.

1.3 Medium-term UK land use change

Afforestation of upland areas was the largest single land use change in Britain in the twentieth century and currently represents perhaps the major medium-term (5-50 years) anthropogenic influence on the British landscape. At its peak in the early 1970s 40,000 ha yr^{-1} of British uplands were being afforested (Robinson and Blyth, 1982). Historical patterns of afforestation in England, Wales and Scotland have varied slightly with peak planting rates occurring in England in the period 1955-1959 (12,000 ha yr^{-1}), Wales in the period 1960-1964 (6000 ha yr^{-1}) and Scotland in the period 1970-1974 (35,000 ha yr^{-1}) (Omerod and Edwards, 1985).

Afforestation of British uplands involves the ploughing of furrows up and down slopes to drain the surface and establish turves into which young trees are planted. The downslope ends of the furrows are connected to deeper drains which normally connect to the nearest watercourse (Stott, 1997(a)). It is now commonly accepted that incision of these deeper drains has the potential to increase bed load supply to the watercourses into which they drain. Estimates range from a near four-fold increase to a more than seventeen-fold increase (Newson, 1980). This increase in coarse sediment production in the upland catchment has strong implications for the channel downstream, particularly in Schumm's deposition zone where increased storage of bed load might lead to lowland channel instability. Indeed, the potential for downstream instability following catchment afforestation has been recognised for years (Tuckfield, 1980; Newson, 1986; Newson and Leeks, 1987), yet to date few studies exist in the literature which specifically examine the extent to which UK lowland channel stability has been affected by upland catchment afforestation.

The reason for this gap in the literature is simple. The transfer time for upland catchment bed load to be transported through the transfer zone to the zone of deposition may be several decades. With the peak rate of afforestation occurring less than 30 years ago, it is only now that lowland channels may be expected to respond.

Upland afforestation also affects the magnitude and frequency of peak discharges at certain points in the forest rotation (i.e. the cycle of planting, growth phase and felling). Increases of up to 40 % in peak discharges have been recorded (Robinson, 1980) and some reduction in low flows has also been noted (Law, 1956). Unlike bed load yields, which may have a long transfer time to the zone of deposition, the impact of altered streamflow on lowland channels may be apparent quickly with a relatively fast morphologic response. Consequently, a further man-induced source for channel response exists where the change may be driven by altered streamflow conditions.

The morphological implications of upland catchment afforestation on lowland channel stability is therefore timely. In the UK the majority of lowland river channels are managed in some way (Sear et al., 1995) and increasingly emphasis is being placed on sustainable management strategies. If the management of lowland rivers whose upland catchments have been afforested is to be sustainable then the impact of any elevated upland catchment bed load yields and altered streamflow conditions on the lowland channel morphology must firstly be understood and secondly built into the management scheme.

1.4 The scope of this project and the thesis structure

This project will examine the medium-term links between upland catchment afforestation and lowland channel instability in a currently destabilised fluvial system containing significant upland catchment forestry, and identify the important factors which have determined the magnitude and locations of the lowland channel responses. The study can be split into four main parts:

1. Chapters 2 and 3

   Firstly, the sedimentologic and hydrologic impact of upland catchment afforestation on the fluvial system is reviewed and the implications for medium-term lowland channel morphology are discussed. The case study catchment is introduced together with the circumstantial evidence for a link between its lowland river channel instability and the upland catchment afforestation. In light of this review the detailed aims and objectives of the study are presented.

2. Chapters 4, 5 and 6

   Secondly, the medium-term history of the study river is examined. The historical investigations aim to determine (1) the medium-term impact of the upland catchment forestry on the upland catchment bed load yields (2) temporal variation in the medium-term patterns of flow in the lowland channel and (3) the spatial and temporal nature of medium-term channel changes both in the middle and lowland reaches of the river.

3. Chapter 7

   Thirdly, the magnitudes and mechanisms of contemporary lowland channel changes are investigated with the aim of identifying the main factors influencing the stability of the lowland channel.

4. Chapters 8 and 9

   Finally, the historical and contemporary data are discussed in the context of the evidence they provide for and against a causal link between the lowland channel instability and the impacts of the afforestation of the upland catchment. The findings of the project are discussed more broadly in terms of the implications for lowland river management and a conceptual model which explains the main findings of the research is presented. The success of the approach taken is discussed and the project conclusions are stated.

2. UPLAND CATCHMENT FORESTRY - HYDROLOGIC AND SEDIMENTOLOGIC IMPLICATIONS FOR LOWLAND CHANNELS

Chapter structure

This chapter is divided into two parts. In part 1 the effect of forestry in upland river catchments is discussed in terms of its impact on bed load yields and runoff to rivers draining them. The major sources and general temporal pattern of bed load from forested upland catchments are examined. Considerable variability in the magnitude of these yields is shown to exist between studies of forested catchments from various geographical locations and with variable catchment physiography. Subsequently the impact of catchment forestry on runoff is reviewed.

In part 2 the potential responses of lowland river channels to modified bed load yields and runoff are considered. Particular attention is given to the role of sediment waves in channel change. Likely medium-term impacts of sediment yield and runoff variation on the lowland fluvial system are examined. Where possible this is achieved through review of direct field studies from catchments containing forestry. More commonly, however, lowland channel response to variations in sediment and water yields is discussed in light of evidence from non-forested catchments, in which measured sediment yield and runoff exhibit patterns similar to those identified from upland catchment studies.

Finally, the research problem with which this study is concerned is stated and the selected field study site is introduced.

PART ONE

THE IMPACT OF FORESTRY ON UPLAND BED LOAD AND RUNOFF

2.1 Upland forestry: impact on bed load yields

The stability of British hillslopes and their regoliths can be extremely variable. Where they are stable, their stability is due largely to the protective characteristics of upland vegetation. However, where surface vegetation is disrupted and channeled overland flow develops, there is the potential to mobilize large quantities of poorly cohesive glacial and glaciofluvial deposits which underlie the surface soils and peat over much of upland Britain. As a consequence, suspended sediment and bed load yields may become elevated where yield is defined as the total bed load outflow from a basin over a specified time period. Such disruption and channeling is a common result of ground preparation for upland forestry, where drainage ditches are often required to lower the water table and thereby reduce the water content of the upland soils and peat prior to tree planting. Furthermore, during and following the establishment of plantations, the drainage ditch networks are maintained to ensure adequate drainage throughout the forest cycle. Ditch erosion, particularly incision, may then provide further localized sources of poorly cohesive material over the life of the forest.

There are several examples of UK studies which have specifically investigated bed load yields from afforested upland catchments, via comparison with adjacent grassland catchment yields (table 2.1). The results show that bed load yields from forested catchments are always greater than from neighbouring grassland catchments.

Table 2.1 Examples of UK paired catchments and bed load yield ratio between forested and grassland catchments.

Author / date	**Catchments studied** **(*F*) = forested** **(*G*) = grassland**	**Bed load yield ratio**
Newson (1980)	Tanllwyth (*F*) and Cyff (*G*), mid-Wales	3.7:1
Newson (1980)	Marchant (*F*) and Conwy (*G*), mid-Wales	17.6:1
Moore and Newson (1986)	Tanllwyth (*F*) and Cyff (*G*), mid-Wales	6.0:1
Johnson (1993)	Balquhidder catchments (*F*) & (*G*), Scotland	7.3:1

Whilst there is little argument that bed load yields are elevated during and following catchment afforestation, the magnitude of the elevation is seen to be highly variable. This is unsurprising given that forest sources of bed load are variable between catchments, and are highly dependent on the catchment geology, physiography and the nature of the forest practices employed. It is also strongly influenced by whether the forestry operation involves both ground preparation and planting (standard in the UK), or is simply the logging of virgin forest (common in studies from the northwest USA). The main forest bed load sources can be summarised as:

1. Forest roads:

Fredriksen, 1970; Madej, 1982; Reid and Dunne, 1984; Duck, 1985; Fukishima, 1987; Madej and Ozaki, 1996.

2. Drainage furrows and ditches:
Newson, 1980; Arkell, 1983; Moore and Newson, 1986; Ferguson and Stott, 1987; Flintham, 1988; Francis and Taylor, 1989; Leeks, 1992; Stott, 1997(a); Stott, 1999.

3. Forest stream channels:
Murgatroyd and Ternan, 1983; Stott, 1997(b).

The following paragraphs aim to highlight the relative importance of the above sources of bed load, particularly in UK forested catchments, and to consider the forest rotation as a temporal control of bed load yields.

2.1.1. Bed load input from forest roads

Examples of literature concerning forestry impacts on bed load yield in the USA are concentrated on the logging of virgin forestry in the northwestern states (e.g. Fredriksen, 1970; Brown and Krygier, 1971; Madej and Ozaki, 1996). In such forestry operations there is no requirement for the ground preparation and ditching which is standard in the afforestation of UK upland catchments. Instead, the major ground disturbance results from the establishment of forest roads. Two main sets of processes have been identified by which forest roads may influence stream sediment yields

1. The production of catastrophic events, e.g. landslides:
Fredriksen (1970) demonstrates that three major landslides, associated with road constructions, were responsible for 93 % of the catastrophic catchment erosion triggered by a single storm in 1964 in three experimental catchments in Oregon. Indeed, in an earlier study in 1961 he was able to demonstrate that approximately 5000 m^3 of debris was delivered to a stream channel by a landslide caused as a direct result of road construction. However, these studies were undertaken in a catchment with relatively steep topography (mean slope gradient of 0.09), which was prone to landslides. As such, landslides are probably of less importance in less steep catchments.

2. Non-catastrophic events, e.g road ditch scour, gullying, road surface sheetwash and overland transport:

Madej (1978) was able to directly quantify sediment yielded from road ditch scour and sheetwash in the Big Beef Creek catchment (32 km^2), California. The study estimates that forest roads contributed an extra 2890 t yr^{-1} of sediment to the creek, with 2450 t yr^{-1} resulting from sheetwash erosion of roadbanks and road beds and 440 t yr^{-1} being a result of road side ditch scour. Indeed, sediment sourced from forest roads is shown to account for nearly half the Big Beef Creek catchment sediment yield. The majority of this sediment (approximately 75 %) however, enters the channel as suspended sediment.

Madej and Ozaki (1996) suggest elevated sediment supply to Redwood Creek, California, to be almost entirely from roads as a result of surface erosion, gullying, road crossings and road fill failures, although no direct evidence of these processes is offered. The study is, however, interesting as rather than solely estimating sediment yields, it provides field evidence of the impact of the elevated sediment supply on the river channel morphology both within and downstream of the forest. Repeat cross profile surveys show the elevated sediment supply to be expressed in the channel as a wave of bed load being transported downstream and causing significant adjustment of the channel morphology. In particular, maximum bed elevation increased by up to 1.5 m and decreased exponentially downstream as the wave attenuated.

Attention has also been given to sediment production from UK forest roads. Duck (1985) investigated the effect of road construction on sediment deposition in loch Earn, Scotland. Deposition of sediment in the Loch following road construction is shown to increase by 20 times with 1824 t of sediment being deposited in the loch over a two month period. In Kirkton Glen, Scotland, the impact of forest road regrading on bed load yield in the first year of clearfelling was also thought to be important (Ferguson and Stott, 1987), possibly providing more sediment than the clearfelled slopes. However, the sediment yields from the roads were not quantified due to a 'lack of meaningful data'. Subsequent work in the same catchment is, however, not conclusive. Stott (1997a) states that, 'the impact of road regrading on bed load yields is small, though it no doubt contributed to the greatly increased suspended sediment loads reported by Johnson, (1993)'

The idea that bed load yields from forest roads can be important during felling and initial afforestation (Ferguson and Stott, 1987) is also echoed by Leeks (1992) in a paper concerned with the temporal pattern of forest sediment yields in the Cwm catchment (1.2 km^2), mid-Wales. Bed load yields were seen to nearly triple, from 12 m^3 yr^{-1} in the year of road construction (1984), to 34 m^3 yr^{-1} in the following year. Peak bed load yields of 55 m^3 yr^{-1} were attained in 1987, four years after road construction. However, the enhanced yields were not sustained following this period inferring that forest road construction has a significant, but short lived impact.

2.1.1.1. Discussion

There is little doubt that suspended sediment concentrations can be enormously affected by forest roads due to associated catastrophic failures and to processes such as sheetwash. It is, though, difficult in many cases to envisage how, over the long-term, bed load yields could be greatly affected unless roads are constructed in close proximity to pathways capable of transporting coarse sediments from the roadway to the fluvial system (i.e. crossing many tributaries). There is evidence from the Cwm catchment , Wales, that during road construction bed load yields may be enhanced for short periods and that this may be an important source of bed load to streams during the establishment of the forest. However, following construction and stabilization of the road yields fall again. Therefore, in the UK, considerably more attention has been given to the effects of ground preparation and drainage ditch erosion because these two sources have the potential to continuously deliver large volumes of bed load as well as suspended sediment to the catchment.

2.1.2 Bed load inputs from drainage ditches

There are several studies which demonstrate a substantial rise in suspended sediment yields as a result of forest furrow and ditch erosion (e.g. Robinson and Blyth, 1982; Soutar, 1989; Francis and Taylor, 1989) but bed load yields have been less widely considered (Stott, 1997a). Literature from the USA, concerned with virgin forestry, contains no data regarding bed load yields from drainage ditches. Indeed, all of the studies concerned with ditches are UK based, the majority being from mid-Wales, and Scotland. Both data sets show the same general trends but the magnitude of response is highly varied, even in similar geographical areas. This reflects the importance of underlying geology and catchment physiography as major controls on bed load yield. Here Welsh and Scottish data sets are examined independently. The greater attention has been paid to Welsh studies which have more extensive data sets available, and in which forest catchment bed load yields are commonly greater than in Scotland. This higher bed load delivery is due to the unusually friable Ordovician and Silurian rock found in mid-Wales which provides abundant gravel-sized particles to the regolith and till from which bed load is commonly yielded (Ferguson and Stott, 1987).

2.1.2.1 Welsh data sets

The Welsh data sets collectively represent the most comprehensive study of bed load yields from forest drainage ditches in the world. Much of the work originates from the Institute of Hydrology and their experimental catchments at Plynlimon and adopts a paired catchment approach. Working in the Plynlimon catchments, Newson (1980) provides an early data set composed of five and a half years of trapped sediment data spanning 1973 - 1978. The catchment bed load yields of two neighboring Welsh upland catchments, the Cyff (3.13 km^2, unditched) and the Tanllwyth (0.89 km^2, ditched), are compared. The ratio of trapped bed load from the Tanllwyth and Cyff is shown to be 3.7:1 over a period of comparable rainfall and runoff. Mean annual bed load yield for the Tanllwyth is calculated as 41.29 t km^2 yr^{-1}. Importantly, Newson is also able to demonstrate that the difference in the ratio of bed load yields is not purely a result of the difference in catchment physiography; the Tanllwyth being smaller and having a higher channel gradient than the Cyff. He uses the empirical formulae of Maner (1958) and Roehl (1962) to calculate the sediment delivery rates (the percentage of annual erosion, per unit area, delivered to the measuring point) for the two catchments. His results show suspended sediment delivery ratios between the Tanllwyth and Cyff in the range 1.2:1 and 1.4:1; much lower than the observed bed load delivery ratio of 3.7:1.

Newson, (1980) was also able to show that the process of ditch incision is real, providing direct measurement of ditch erosion via repeat cross profiles (figure 2.1). It is clear from these repeat cross profiles that drainage ditches in mid-Wales can be extremely active. Commonly, ditches eroded by up to 40% over the 26 year study period, with over 75% erosion not being uncommon. Whilst the changes to cross section morphology were varied, the majority of sediment yielded was seen to be a product of ditch incision. Total erosion depth was limited by the depth of deposits overlying the bedrock. It should be noted that erosion in the Tanllwyth ditches is not necessarily representative of all forest ditches in mid-Wales, however further direct measurements for comparison have not been published.

The long periods between bed load trap emptying for the Cyff and Tanllwyth reported by Newson (1980) prevented investigation of the temporal patterns and volumes of coarse sediment production from ditches and transport to the nearest stream with respect to patterns of rainfall and ditch discharges. High temporal detail is important in the understanding of patterns and magnitudes of bed load delivery to piedmont streams and hence the lowland channel responses. Consequently, further smaller traps were installed in both the Tanllwyth and Cyff catchments, in order to investigate the temporal patterns and magnitudes of ditch erosion, storage and sediment transport (Arkell, 1983). The results from these traps show short-term yield from the Tanllwyth ditches to be highly variable, being between 3.5 and 20.5 times those of the Cyff. The results strongly suggest that supply limitation is an important temporal control of bed load delivery from ditches to upland streams.

A later study from the Tanllwyth and Cyff catchments based on a more extensive twelve year bed load yield record (Moore and Newson, 1986) sees the average ratio

Figure 2.1 Examples of forest ditch erosion (shaded areas) over a 26 year period in the Plynlimon Experimental catchments. Note the majority of erosion is of the bed where poorly cohesive regolith material is in abundance. From Newson (1980).

of bed load yield calculated as 6:1; considerably higher than the earlier ratio of 3.7:1. Much of this increase can be attributed to a single storm in August 1977, which affected mainly the Tanllwyth catchment. Following the storm more than 80 t of bed load was collected in the trap; equivalent to almost 30 % of the entire bed load yield between 1973 and 1980. Little bed load was collected in the Cyff. The result serves to further demonstrate the variability of bed load yields between forested and non-forested catchments and the sensitivity of upland catchments to extreme meteorological events, which can be very localised in upland Wales (c.f. Newson, 1979).

Moore and Newson (1986) used stepwise regression analysis to identify the dominant factors influencing bed load transport in the forested and moorland catchments. The results suggest that instantaneous peak discharge and duration above the transport threshold dominate in accounting for the high temporal variation in bed load yield. 61 % of the variance in the bed load yield of the Tanllwyth was shown to be explained by these variables. Again, this result serves to underline the potential of high magnitude storms to significantly elevate bed load yields in upland catchments, and helps to explain why the bed load yield ratios presented in table 2.1 vary to such a high degree.

2.1.2.2 Scottish data sets

Similar data sets are available for the forested Monachyle (7.7 km^2) and Kirkton (6.85 km^2) catchments at Balquhidder in central Scotland (Stott *et al.*, 1986; Stott, 1987; Ferguson and Stott, 1987; Johnson, 1988; Johnson, 1993; Stott, 1997a). The Monachyle catchment was studied during the forest development and ground preparation phase, meaning that the impact of ground preparation on bed load yields was directly monitored. Between 1983 to 1985, when the catchment was undisturbed, and 1986 to 1989, when the ground was ploughed and drained, gross bed load yields were elevated from 2 t to 3 t. At Kirkton Glen the impact of clear-felling on the already maturely forested catchment was examined over the same time periods. Gross bed load yield was elevated from 17 t prior to clear-felling to 19 t after clear-felling. Converted to ratios, bed load yield could be seen to vary with the forestry phase, from 1.3:1 between the undisturbed moorland and ploughed catchment, to 7.3:1 during mature forest and 8.3:1 as a result of clear-felling. Compared to the Tanllwyth catchment which had a mean bed load yield of 41.29 t km^2 yr^{-1} between 1973 and 1980, the mature forest in the Kirkton Glen catchment yielded just 1.23 t km^2 yr^{-1} of bed load. This comparison again serves to demonstrate the variability in bed load yields between forested catchments. More importantly, though, the results serve to demonstrate how the forest rotation is a potentially important control of bed load yield from drainage ditches.

2.1.3 Bed load inputs from stream banks

Sediment supply to stream channels from bank erosion has been shown to be of importance in the overall sediment budgets of some rivers. Hill (1973) states that it is possible for bank erosion to be the major source of sediment to river channels. Indeed, Kirkby (1967) found that 93% of the total sediment removed from the mountain catchment of the Water of Deugh, Scotland, was via the erosion of river bluffs. Yet, despite such examples, there have been relatively few studies which have measured upland stream bank erosion directly (Blacknell, 1981, Stott and Marks, 1998; Stott, 1999).

In afforested catchments recognition of the importance of drainage ditch erosion to sediment yields led to substantial studies. Drainage ditches are, in effect, small, artificial stream channels and it is therefore surprising that relatively little attention has been paid to evaluating the impact of catchment afforestation on forest stream bank erosion processes, in particular, the potential of elevated bank erosion rates to increase forested catchment sediment yields. Where comparison of stream bank erosion rates within forested and moorland streams have been undertaken, the results have not been conclusive (Murgatroyd and Ternan, 1983 and Stott, 1997b).

Murgatroyd and Ternan (1983) compare rates of bank erosion within forested and non forested reaches of the Narrator Brook, Dartmoor. Mean bank erosion is shown to have increased from 0.7 mm yr^{-1} to 5.2 mm yr^{-1} following reach afforestation. Bed aggradation within the forested reach is 3.7 times that outside the forest. Active bank erosion is largely attributed to the suppression by the forest of a thick grass turf, and its associated dense network of fine roots which act to enhance bank stability. Additionally, log jams and debris jams in the stream channel were considered important due to their potential to lead to flow deflection and enhanced rates of bank erosion (e.g. Davis and Gregory, 1994).

When compared to other published stream bank erosion rates (table 2.2) it is apparent that the erosion rates from Narrator Brook are an order of magnitude lower than those of many forest streams of similar size suggesting an unusually stable channel.

Table 2.2 Comparison of mean annual forest stream bank erosion rates within the UK.

Author / date	**Study site**	**Mean annual bank erosion (mm yr^{-1})**
Murgatroyd and Ternan (1983)	Narrator Brook, Dartmoor, Devon	5.2
Lewin et al. (1974)	Maesnant, mid-Wales	30
Hill (1973)	Clady and Crawfrodsburn, N. Ireland	30-66
Davis and Gregory (1994)	New Forest, Hampshire	63.5
Stott (1997b)	Kirkton Glen, Scotland	47
Stott and Marks (1998)	Tanllwyth, mid -Wales	64.3

Murgatroyd and Ternan's result that the forested reaches experience higher rates of erosion than non-forested reaches is also contradicted by Stott (1997b) The findings of the two studies are compared in table 2.3. Stott (1997b) shows a decrease in stream bank erosion rates following catchment forestry which he attributes to the restriction of frost in banks under forest canopies, where the mean air temperature range is 26 °C in the Monachyle catchment and only 12°C in Kirkton Glen.

Table 2.3 Stream bank erosion in forested and non-forested channels. Comparison of Murgatroyd and Ternan (1983), Stott (1997b) and Stott (1999).

Author / date	**Catchment / reach vegetation**	**Erosion rate (mm yr^{-1})**
Murgatroyd and Ternan (1983)	Narrator Brook, non-forested	0.7
	Narrator Brook, forested	5.2
Stott (1997b)	Monachyle catchment, non-forested	59
	Kirkton Glen catchment, forested	47
Stott (1999)	Cyff, non-forested	65
	Tanllwyth, forested	70 (increasing to 95 post-felling)

2.1.3.1. The role of frost

Erosion pin measurements in the Monachyle and Kirkton Glen catchments (Stott, 1997b) demonstrate lower mean bank erosion rates for the forested Kirkton catchment (47 mm yr^{-1}) compared to Monachyle (59 mm yr^{-1}). Correlation coefficients for mean daily bank erosion rates and various stream flow and frost indices demonstrate clearly that frost indicies correlate better with erosion rates than any of the stream flow indicies. The reduced bank erosion rate in the forested catchment is attributed to higher stream bank temperatures in forested catchments. Bank temperatures in Kirkton are on average 3.7 degrees higher than those of Monachyle reducing the incidence of needle ice within the banks.

The influence of frost and seasonality demonstrated by Stott, (1997b) is noted in other studies (e.g. Wolman, 1959; Lawler, 1986; Lawler, 1993b), and has been shown to be of importance in the Welsh Upland catchments. The formation of needle ice within banks forces sediment away from the bank both on the surface and within growing needles. On melting this material may fall to the foot of the bank or remain as a loosely bound layer easily washed away by the next rise in stage. Such a process undoubtedly plays an important role in the production of suspended sediment. It is less probable that it is responsible for elevated bed load yields.

2.2 Bed load yields and the forest rotation

The literature reviewed previously offers considerable insight into the individual sources and mechanisms by which mature forest may deliver bed load to stream channels at a given point in the forest rotation. However, few of the studies examine these sources and mechanisms in terms of their variability throughout an entire forest rotation. This is perhaps not surprising given that the forest rotation, from planting to felling, may take in excess of 50 years. To study the impact of the rotation on bed load yields in a single catchment would therefore require an extremely long study.

The differing forest phases monitored in the Scottish Monachyle and Kirkton Glen catchments do, however, provide an insight into the difference in bed load yield during the transition between moorland and ground preparation, and mature forest and clear-felling. In addition, Leeks (1992) provides a summary of the impact on bed load yields in several Welsh catchment studies of ground preparation, planting, mature forest and clear-felling.

Both the Scottish and Welsh catchments show an increase in bed load yield as a result of ground preparation, however, the sharp, lagged, short-term rise in bed load yield following ditching, furrowing and, in particular, road development in the Llanbrynmair catchment (mid-Wales) is less well defined in the Scottish catchments. During the period of mature forest canopy, bed load yield in both the Scottish and Welsh catchments is artificially elevated above yields in neighbouring, or previously moorland catchments. Felling in Kirkton Glen is followed by a slight increase in bed load yields (2.5 t km^2 yr^{-1} compared to 2.2 t km^2 yr^{-1}) however, results from the Hore catchment in mid-Wales (Leeks, 1992) show a fall in bed load yield from 11.8 t km^2

yr^{-1} to 8.8 t km^2 yr^{-1} directly after felling. This decrease is attributed to the impact of within-ditch woody debris jams which act to restrict transport of bed load through the ditches to the stream channel. Following the removal of the jams bed load yields responded similarly to those in Kirkton Glen by increasing to 54.5 t km^2 yr^{-1}.

As a result of the evidence provided by the Welsh catchment studies, Leeks (1992) presents a model of the impact of the forest rotation on upland stream bed load yields. The model is presented in figure 2.2. Bed load yields experience two peaks which follow site preparation and felling. The second peak is preceded by a drop in yield caused by brashings in the ditches protecting the bed and reducing bed load transport. The peak following felling is less significant in terms of stream sediment loads due to the already elevated bed load yields which precede it. However, the peak at site preparation and subsequent maintenance of elevated bed load yield may have significant morphological impact on forest streams. Following downstream transport, lowland channel morphology may also by influenced, particularly if the increased bed load yield is expressed as a sediment wave, as is the case in the study of Madej and Ozaki (1996).

2.3 Sediment delivery ratios

The studies presented in this review highlight the variability in bed load yields per unit area, per unit time, when comparing different catchments. Whilst some of this variability may be accounted for by sampling differences and variability in rainfall / runoff processes within the catchments, much of it is a result of the physiographic and geological constraints of the catchments to yield sediment.

Maner (1958) and Roehl (1962) independently provide a method by which the sediment delivery ratios of two catchments, can be compared using the following empirical formulae based on a variable which describes catchment physiography. The sediment delivery ratio is the ratio between the sediment yield at a specific point in the catchment and the total sediment entering the catchment upstream of that point.

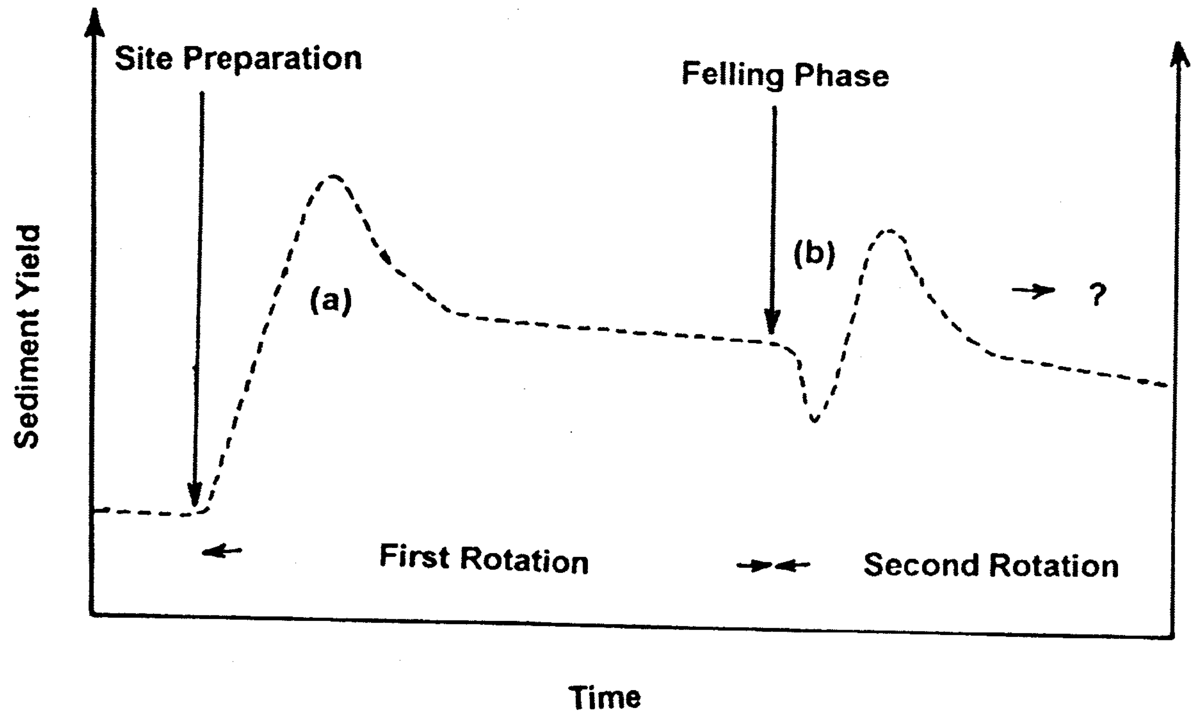


Figure 2.2 Sediment yields and the forest rotation. An initial peak in sediment yield (a) follows site Preparation before yields slowly decline subsequently. Following felling a drop in yield (b) due to woody debris clogging drainage ditches preceeds another peak. After Leeks (1992).

Maner (1958): $\log DR_e = 2.94259 - 0.82363 \text{ colog } R/L$ (equation 2.1)

Roehl (1962): $\log DR_e = 2.88753 - 0.83291 \text{ colog } R/L$ (equation 2.2)

Where:

DR_e is the estimated sediment delivery rate (the ratio between the annual gross erosion in the catchment and the annual sediment yield)

R/L is the catchment relief to length ratio (descriptive of catchment physiography)

The formulae they derive are extremely similar and explain 97 % and 88 % of the sediment delivered to the USA catchments respectively. They also serve to demonstrate the important control the catchment physiography has over sediment delivery.

As previously mentioned, Maner and Roehl's formulae, mainly relating to suspended sediments, have been used by others (e.g. Newson, 1980) to check that the higher bed load yields experienced in forested catchments are not a result of physiographic differences in the paired catchments. Further to this, Stott (1997A) was able to show that in catchment comparisons the % of catchment area forested is important. By combining the bed load yield and forest cover records of Scottish and Welsh catchments he was able to demonstrate the existence of a weak positive exponential relationship between percentage forest cover and bed load yield. The relationship is presented in figure 2.3. However, Stott's relationship takes no account of the variable catchment geology, physiography, pedology or climate, which probably explains why the relationship has a high residual scatter.

2.4 Summary

The literature is in common agreement that afforestation of UK upland catchments leads to elevated bed load yields within the streams draining them.

Studies from the North West USA have strong emphasis on forest roads as a major source of enhanced bed load yields. This is unsurprising because forestry operations involve only logging of virgin forestry. Consequently there is little ground disturbance away from roadways and none of the ditching and furrowing associated with commercial forestry in the UK. In the UK road density is often low and forest roads tend to be long term, relatively stable features of the forest. Commonly they are located several hundred metres from the stream channel and hence have a low influence on forest stream bed load yields. Consequently, relatively little attention has been given to road generated bed load. Where studied, short-term bed load yields at the period of road construction may be considerably enhanced. However, the longer-term enhancement of bed load yields from forest roads is thought to be of minor importance compared to those from forest ditches, although they can be a major source of suspended sediment.

The literature concerned with sediment yields from forested stream bank erosion is difficult to interpret and contains some apparent contradictions in results. However, it is clear that frost plays an important role in stream bank erosion providing an annual temporal control on yields. Due to the short lengths of stream channel within a forested catchment when compared to the combined lengths of furrows and ditches it seems reasonable to consider additions to catchment bed load yields from stream bank erosion to be of low significance.

There are now several studies in the literature, and several UK based, long term data sets available, concerned with bed load yield from forest ditches. These data sets provide strong evidence that drainage ditches have the potential to deliver large volumes of bed load to afforested catchment streams over long periods. Following transport to the main channel there is little doubt that upland stream bed load yields are greatly enhanced as a result of drainage ditch erosion.

Bed load yields from forests are shown to be highly variable and dependent on precipitation and forest hydrology, particularly the magnitude and duration of storms, the catchment size and topography, drainage density and the underlying geology. Yields are also shown to vary within the forest rotation.

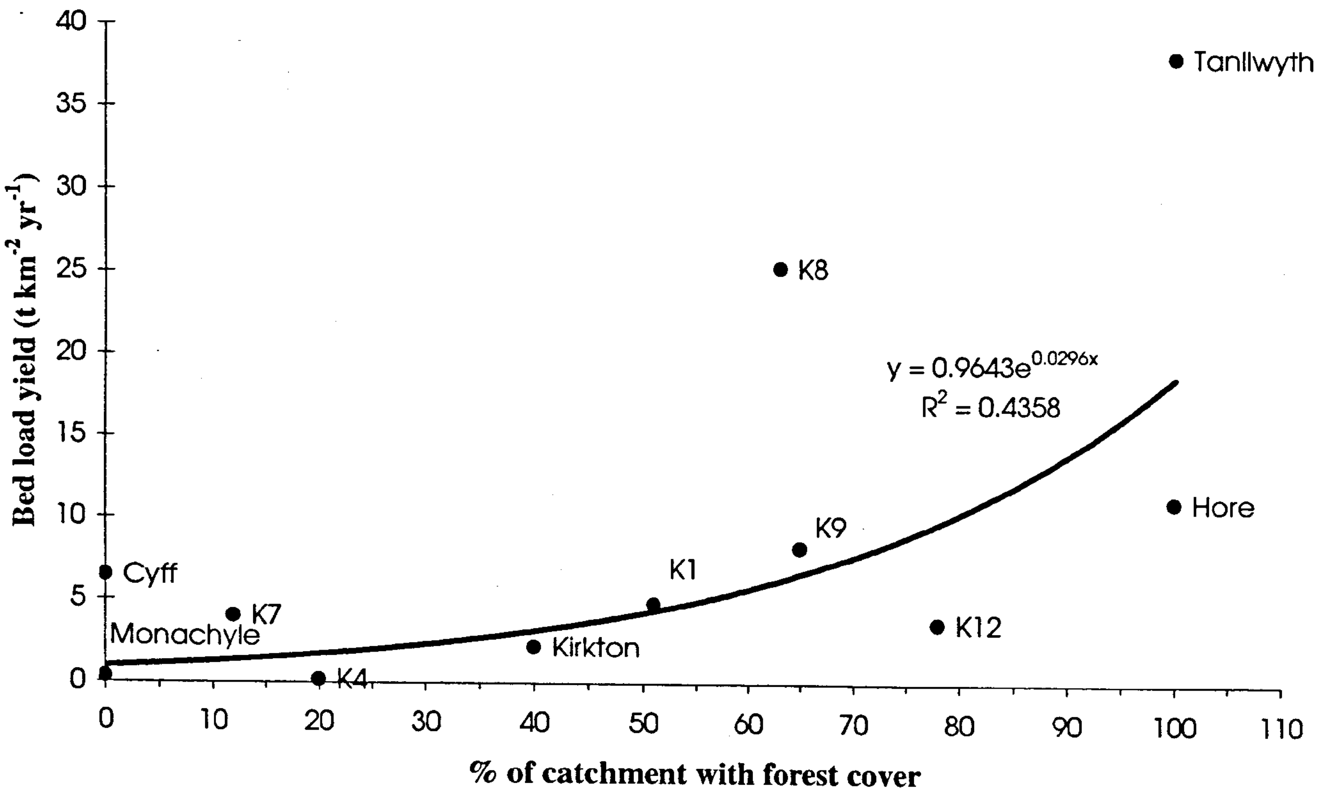


Figure 2.3 The effect of catchment % forest cover on bed load yields in UK catchments. Scatter plot of subcatchment bed load yield versus % catchment forest cover. Graph shows data from Kirkton Glen, Balquhidder, Scotland with the Kirkton (K1-K12) and Monachyle yields (Stott, 1987 and Johnson, 1993) and the Hore, Wales (Leeks, 1992). The Cyff and Tanllwyth catchments at Plynlimon, mid-Wales (Moore and Newson, 1986) are also shown. From Stott (1997a).

2.4.1 Implications for lowland channel stability

The increased quantity of bed load released as a result of catchment afforestation will eventually enter the main river channel where it will undergo transport to lowland reaches. Here it is possible that the channel will respond via aggradation and that some of the sediment will enter storage within bar forms, particularly if a mechanism exists by which the forest-derived bed load can be deposited in restricted storage reaches. In such a scenario, channel quasi-equilibrium will be disrupted and the lowland channel may become destabilized. The ways in which this may occur are discussed later.

2.5 Impact of upland afforestation on runoff regime

2.5.1 Introduction

Land drained prior to and following the establishment of forestry in the UK will have an altered runoff regime. Runoff becomes quickly channelled and natural vegetation is replaced by a coniferous canopy altering considerably interception and evapotranspiration, thereby affecting runoff from the catchment. The impact of this on river flows, particularly peak flow events capable of achieving channel morphologic adjustment, is of considerable importance to the stream channels draining the affected catchment both within the afforested reaches and in lowland reaches. River channels will adjust to these altered hydrologic inputs. Alterations to river sediment yields require a period for sediment to be transported and deposited before lowland reaches are seriously affected. In contrast, modified hydrologic regime will have a more immediate impact on the geometry of the lowland channel.

Interest has been paid to the impacts of afforestation on runoff regime in the UK, mainly as a result of concerns over water resources. The following paragraphs review the literature stemming from that interest. In particular the impact of afforestation on total runoff and flood discharges are examined and, as before, the role of the forest rotation is considered.

2.5.2 Impact of afforestation on total runoff

There are several examples of studies concerned with the effect of upland afforestation on runoff which set about to answer the question 'How much water does forestry use?'. The results from three key studies are presented in table 2.4.

Table 2.4. Water loss from forested catchments compared to water loss from open water.

Author / date	**Study location**	**% water loss from forest canopies compared to open water**
Law (1956)	Stocks Reservoir, northwest UK	136 %
Rutter (1964)	Castricum lysimeters, Netherlands	110 – 120 %
Clarke and McCulloch (1979)	Upper Severn catchments, mid-Wales	128%

The work of Law (1956) on Stocks Reservior on the headwaters of the River Ribble, UK, represents the first British attempt to answer this question. Law (1956) showed a loss of 290 mm yr^{-1} from a 450 m^2 forest lysimeter, which he inferred to be a direct result of the afforestation of the Stocks reservoir catchment in northwest England. This figure represents evaporation for the catchment as being 36 % greater than that from open water. Indeed, Law states firmly that afforestation led to significant costs in replacing lost water resources and that the foresters owed the Water Authority £550 ha^{-1} yr^{-1} in loss of water. Law's experiments, however, were on relatively small plots of land (a natural lysimeter covering only 450 m^2 in a block of woodland covering 2,400 m^2) and hence serious questions about the validity of extrapolating the results to large catchments have been raised.

The results of Rutter (1964), based on data from a lysimeter in the Netherlands, serve to offer support to the patterns of water loss observed by Law. Rutter's evidence suggested that water loss from plantation canopy exceeded that from open-water evaporation by 10-20%, a value up to 26 % lower than that reported by Law. He concluded that the main causes for the higher values of actual evaporation from forests were the darker colour of the vegetation, especially of conifers, leading to greater adsorption of energy and the greater aerodynamic roughness. In comparison, Clarke and McCulloch (1979), show evaporation rates in the upper Severn catchment, UK, to be 28 % greater than that of open water – a figure roughly comparable to that given by Law. Although there is variability in the total loss, all of the above studies serve to demonstrate that forest canopies have the potential to lose a great deal of water.

More recently the experimental catchments of the Wye (1055 ha) and Severn (870 ha) in mid-Wales have been used to examine the same fundamental question but experimentation has been developed on a far greater temporal and spatial scale. Texts relating to the methodologies employed and the preliminary results from the Plynlimon catchments have been published at stages throughout the duration of the experimentation (i.e. Newson, 1976; Clarke and McCulloch, 1979; Newson, 1979; Kirby et al., 1991) A summary of the results of the full eleven year study are reported in Kirby et al. (1991) and show a total loss from the forested canopy of nearly double that of open water. This translated to a decrease in stream runoff of 15%.

2.5.3 Impact of afforestation on peak discharge

Whilst an overall, long term, estimate of loss of runoff from catchment afforestation is of interest, particularly to the water resource industry, it is the temporal integration of that loss, particularly its effect on the flood hydrograph, which is of far greater interest to this research. This is because the high rainfall, steep channel gradients and the relatively low levels of soil moisture deficit which build up over the summer within upland catchments make them an important source of flood runoff (Kirby et al., 1991). It is also important to quantify altered runoff in terms of the forest rotation. It would seem probable that the flood hydrograph of a stream draining a felled or recently planted catchment, with little vegetation cover and no mature canopy, will be substantially different to that of a stream draining mature forest.

The alteration of the flood hydrograph following the ground preparation and planting phases of forestry is reported in Robinson (1980) for the Coalburn Catchment, Cumbria, UK and in Leeks and Roberts (1987) for the Cwm Catchment, mid-Wales. Using one hour unit hydrographs (NERC Flood Studies Report, 1975) Robinson (1980) was able to demonstrate an approximate halving of time to peak discharge and an increase of about 40% in peak discharge (figure 2.4) following pre-afforestation ditching of the Coalburn catchment. Total storm runoff as a percentage of total storm rainfall was also analysed at the Coalburn catchment. The relationship between rainfall and runoff was much stronger after drainage possibly indicating a decrease in the importance of soil antecedent moisture conditions. Results from unit hydrograph analysis at the Cwm catchment (Leeks and Roberts, 1987) support the Coalburn results. Drainage was shown to significantly increase peak flows, the magnitude of the increase being dependent on the duration of the storm event. A 10 hour duration rainfall produced an increase in peak discharge of 5% whilst a 20% increase in peak discharge is reported for a storm of 5 hours duration.

The Severn experimental catchments at Plynlimon were operated with mature coniferous canopy cover throughout the experimental period (Leeks and Roberts, 1987; Kirby et al., 1991). As a consequence the data offer no information about water yields during the ground preparation, planting or clear felling phases of the forest rotation, arguably the phases when the impact of land use change is likely to be highest. However, it does provide detailed analysis for mature forest. A sample of 18 paired flood events (i.e. events which occurred in both the forested and grassland catchments), over an eleven year period, mostly smaller than the mean annual flood, were analysed. The data are presented in table 2.5.

Table 2.5 Comparison of mean values of hydrograph parameters in the grassland Wye and mature canopy Severn upland catchments.

	Severn (forested)	**Wye (grassland)**
Number of events	18	18
Peak flow (cumecs)	8.46	11.80
Peak flow (cumecs km^{-2})	0.97	1.11
Percentage runoff	45.5	51.4
Lag (hours)	3.2	3.2

When adjusted for catchment area, the grassland Wye can be seen to have slightly higher peak flows than the Severn. It is also seen to have a greater percentage runoff, a result which confirms the findings of Clarke and McCulloch (1979). Lag time is shown to remain unaffected. The results suggest that for small storms, the effect of a more intensive forest drainage network is compensated for by the effects of forest canopy interception. For greater magnitude events, the forest drainage network is unlikely to have a greater efficiency at routing water than natural runoff and soil pipe mechanisms. Consequently there is little observed difference in the flood hydrographs of the two catchments.

Hudson and Gilman (1993) also analysed the patterns of stream flow from the Severn catchment over a longer study period (1969 – 1988). Whilst their results do not dispute the findings presented above, they do show a pattern of increasing stream flow over the study period which can not be explained by increased rainfall. They interpret their findings as being a result of a decline in forest canopy evapo-transpiration with increasing canopy maturity.

2.5.4 *Summary*

2.5.4.1 *Runoff during mature canopy phase*

During mature vegetation, which represents the majority of the forest rotation, the evaoptranspiration from a forest canopy commonly exceeds that of open water. As a consequence total runoff in afforested catchments are reduced. In Wales the reduction is estimated at 15% and results mainly in decreased low flows. There is evidence to suggest that the magnitude of this decrease is reduced with increasing canopy age. Following small storms, peak discharge is slightly reduced and following large storms peak discharge remains unchanged

2.5.4.2 *Runoff during ground preparation and immature canopy phase*

During the short period of time during each forest rotation when canopy is removed and ditches are exposed the hydrologic response is very different. Flood peaks have been shown to increase by 40% and the flashiness of the hydrograph increases. The duration of rainfall has also been shown to be an important control on flood peaks with shorter storms producing a greater comparative increase in peak discharge. Clearly, these results have implications for lowland fluvial stability of streams and rivers draining afforested upland catchments with channel stability being threatened during periods of ground preparation and immature canopy.

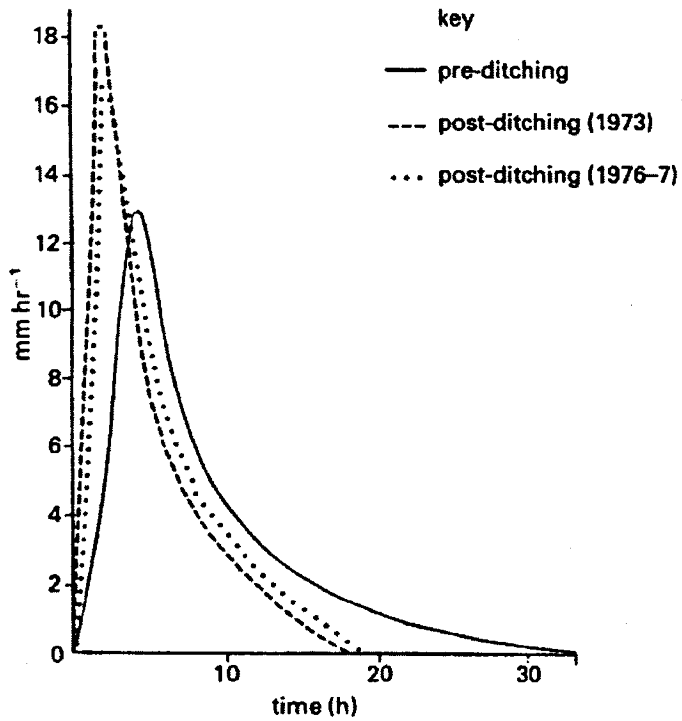


Figure 2.4 Unit hydrograph response to forest ditching in the Coalburn Catchment between 1973 and 1977. Note a flashier hydrograph with a decreased lag time to peak discharge and a reduced recessional limb. After Robinson (1980).

PART TWO

IMPACT OF ELEVATED BED LOAD AND RUNOFF REGIME ON LOWLAND RIVER CHANNEL MORPHOLOGY

2.6 The impact of elevated sediment loads - introduction

Additions to a river's sediment load can be either exogenous or endogenous. Exogenous sediment is that which enters the channel as a result of catchment processes beyond the direct influence of the river channel processes. Such sediment is commonly a result of anthropogenic activity within the catchment, such as the elevated bed load yields associated with upland catchment forestry presented previously, mining, urbanisation or catchment denundation processes (e.g. landslides). Exogenous sediment may be responsible for driving considerable channel change especially if it forms a major addition to the overall bed load yield of a river or forms a sediment addition to a channel already close to its sediment transport capacity. Endogenous sediment is a product of within channel processes – particularly the reworking of bed material and bank erosion and may also form an important addition to the overall sediment load of a river measured at a finite point.

River channels exist to convey exogenous and endogenous sediment as well as the water supplied to them, but are also a product of that conveyance. Consequently, the geometry of a river channel adjusts to the sedimentologic and hydrologic conditions imposed upon it through a complex array of process-response linkages (e.g. Mackin, 1948; Leopold and Maddock, 1953; Leopold and Langbein, 1962). These linkages are shown schematically in figure 2.5 where it can be seen that the sediment load input to a river system, be it from land use, vegetation changes and soils or bank material, is able to influence other areas of the system, in particular bed form geometry and channel width and depth. There are many examples of channel morphology parameters responding to altered loads, for example Orme and Bailey (1971) report a 200 % increase in channel width and a decrease in depth of the Monroe Canyon River, California following sediment load increases. Similarly, O'Loughlin (1970) reports a 40 % increase in stream bed area of streams draining the Reefton experimental catchments in New Zealand. Here, sediment yield was elevated

Figure 2.5 Interrelationships in the fluvial system. Relationships are indicated as direct (+) or inverse (–). Arrows indicate the direction of influence. **From Knighton (1998).**

by small landslides following high magnitude storms in the forested catchments. These reported channel changes are of the type suggested by Werritty and Brazier (1991) who summarise the likely responses of channels to changes in the hydrologic and sediment loads (table 2.6). Because altered loads have the potential to cause significant channel change, much recent attention has been given to identifying and understanding the processes which drive channel responses, particularly in respect of channel responses following increased exogenous sediment loads.

Table 2.6 Channel responses to altered sediment and hydrologic loads. After Werritty and Brazier (1991).

Change in water (Qw) and sediment (Qs) discharge	Impact on river channel
Qs+ Qw=	Aggradation: channel instability, channel widens and shallows
Qs- Qw=	Incision: channel instability, channel narrows and deepens
Qw+ Qs=	Incision: channel instability, channel widens and deepens
Qw- Qs=	Aggradation: channel instability, channel narrows and shallows
Qs+ Qw-	Aggradation ++
Qs+ Qw+	All morphological activity intensified?
Qs- Qw-	Decreased level of morphological activity
Qs- Qw+	Incision ++: channel instability, channel deepens, width change uncertain

2.7 River channel conveyance of elevated sediment loads – the wave model

In the last century the increased anthropogenic disturbance of river basins, in particular land use change and its influence over sediment yields, has generated considerable interest in the downstream conveyance of sediment, and its impact on channel morphology. This interest can be traced back to the study of Gilbert (1917) in which the impact of mining sediment deposits in the Sierra Nevada were investigated. Gilbert studied the aggradation of the river bed of the Sacramento Valley resulting from the influx of mining sediment indirectly, by measuring the changes in the level of low flow stages at three gauging stations. After mining was prohibited channel beds began to incise and low flow stages decreased. Gilbert inferred from the pattern and timing of the bed elevation changes he observed that sediment loads had increased and decreased accordingly and that the mining sediment was being transported downstream in a wave.

Gilbert recognised the passage of a sediment wave via morphological change, notably a pattern of channel aggradation and degradation which could be closely associated with the cessation of mining activities and which extended over a period much greater than the single flood event. Sediment waves can, however, also be thought of in terms of bed load transport rates where a wave can be recognised by a peak in bed load transport rates which may or may not translate downstream (Benda and Dunne, 1997).

The sediment wave model of Gilbert has had a lasting impact on fluvial geomorphology and a large number of studies exist in the literature in which sediment waves have either been directly observed or inferred from channel changes. A range of these studies are presented in table 2.7. Common morphological changes resulting from the translation of a sediment wave through a reach are:

- General bed aggradation followed by degradation – either as regular or irregular pattern (Gilbert, 1917; Griffiths, 1979; Madej, 1982; Beschta, 1983; Pickup et al., 1983; Meade, 1985; Roberts and Church, 1986; Knighton, 1988; Madej and Ozaki, 1996)
- Increase in channel width (Madej, 1982; Roberts and Church, 1986; Knighton, 1988; Madej and Ozaki, 1996; Jacobson and Bobbitt Gran, 1997)
- Decrease in bed gradient (Madej, 1982, Madej and Ozaki, 1996)

- Increase in gravel bar area (Madej, 1982; Meade, 1985; Wathen and Hoey, 1998; Jacobson and Bobbitt Gran, 1999)
- Increased sinuosity (Madej, 1982)
- Aggradation of pools (Roberts and Church, 1986; Wathen and Hoey, 1998; Wohl and Cenderelli, 2000)
- Zones of decreased stability (Jacobson and Bobbitt Gran, 1999)

It is also clear from the table, which by no means represents an exhaustive list of sediment wave studies, that sediment waves are relatively common fluvial features and vary in magnitude considerably. Indeed examples of sediment input magnitude vary from the extremely large inputs observed in the Sacramento Valley (Gilbert, 1917; James, 1999) and Ringarooma River (Knighton, 1988), of more than 1 billion m^3 in 31 years (323 million m^3 yr^{-1}) and 40 million m^3 respectively (374,000 m^3 yr^{-1}), to relatively small inputs in Big Beef Creek (Madej, 1982) of 4394 m^3 yr^{-1}. Recognising this variability in magnitude, Nicholas et al. (1995), who use the term sediment 'slug' rather than wave, present a hierarchical classification of sediment waves (tables 2.8 and 2.9) based on magnitude. They define a sediment slug as a body of clastic material associated with disequilibrium conditions in fluvial systems over time periods above the event scale and that definition is applied to sediment waves in this study. The Nicholas et al. (1995) sediment slug classification is equivalent to bed form scale classes and infers the morphological consequences of the different wave scales. The classification suggests that whilst a macroslug may be capable of invoking minor channel changes analogous in scale to individual barforms, a megaslug or larger is required to invoke major channel change. The classification also suggests that there is a general increase in the importance of extrinsic factors (meaning catchment sediment supply) compared to intrinsic factors (meaning fluvial entrainment processes) as slug magnitude increases as shown by the dominant controls. The slug categories presented by Nicholas et al. (1995) are not mutually exclusive suggesting that the within-network accumulation of several macroslugs could produce a megaslug and the associated dominant controls could alter as a response.

Table 2.7 Examples of field studies in which channel changes have occurred in response to the passage of a sediment wave.

Author / date	Study location	Sediment source	Morphological impact
Gilbert, 1917	Sacramento Valley, Sierra Nevada, USA	**Exogenous**: mining sediments	The elevation and subsequent incision of the river bed as inferred from variation in the low flow river stage as recorded at three gauging stations.
Griffiths, 1979	Lower Waimakariri River, New Zealand	**Endogenous**: variability in the upstream bed load transport capacity	Sediment translation waves caused by downstream changes in bed load transport capacity resulting from downstream changes in hydraulic geometry. Causes irregular fluctuations in bed level. Sediment translation waves are offered as the mechanism by which bed load is transported in all alluvial channels irrespective of channel pattern provided hydraulic geometry varies downstream.
Madej, 1982	Big Beef Creek, Puget Lowland, Washington, USA	**Exogenous**: sediment supplied as a result of forestry. Sediment yield increased from 22 t km^{-2} to 185 t km^{-2}	Channel aggradation (0.5 – 1 m) in the lower reaches, degradation in the middle reaches (approx. 1.1 m) and little change in the upper reaches. Pattern is consistent with the passage of a sediment wave. A decrease in channel gradient from 0.090 to 0.085 and increase of sinuosity from 1.30 to 1.32 is reported in the lower reaches. Channel width increase and growth of gravel bars is also reported in aggrading reaches.

Author / date	Study location	Sediment source	Morphological impact
Meade, 1985	East Fork River, Wyoming, USA	**Endogenous**: Elevated rates of tributary erosion	Successive pulses of bed load that have the form of composite dune fields moving from storage area to storage area with successive pulses of water discharge
Roberts and Church, 1986	Disturbed watersheds, Canada	**Exogenous**: Clear cut logging and landslide generation	Increases in riparian erosion rates. Channel width increases of between 50 and 190 % reported. Local infilling of pools.
Knighton, 1988	Ringarooma River, Tasmaina	**Exogenous**: 40 million m^3 mining sediment	Successive phases of aggradation and degradation progressing downstream. Aggradation was most rapid close to major supply points. Where valley floor was broad width increased by up to 300 %. After the peak of mining activity beds are incising by up to 0.5 m yr^{-1} and channel pattern has changed from braided to non-braided.

Author / date	Study location	Sediment source	Morphological impact
Madej and Ozaki, 1996	Redwood Creek, Califonia, USA	**Exogenous:** Commercial timber harvesting	Initial increase in channel width (between 150 and 300 %) and decrease in channel gradient. Sediment supply was heavily influenced by rainfall with a single flood in 1964 responsible for a 90 % increase in stored sediment in the upper reach. The amplitude of the sediment wave decreased downstream exponentially ($P=1.6\ e^{-0.16x}$ where P is the peak aggradation and x is the distance downstream).
Lisle et al., 1997	Experimental flume channel	**Exogenous:** Mid-flume input	Wave dispersed upstream and downstream attenuating without translation. This occurred because sediment that was eroded from the introduced wave and deposited downstream was equaled by incoming sediment trapped upstream of the wave. Barforms showed no systematic change in volume, celerity or transport rate.
Wathen and Hoey, 1998	Allt Dubhaig, Scotland	**Endogenous:** Avulsion generated wave	Results suggest that where the sediment wave is deposited does not always correspond with areas of subsequent erosion. Gravel bars aggraded and prograded (in excess of 0.45 m in places) and adjacent pools aggraded in response to wave. Zones of erosion, commonly adjacent to the regions of wave deposition, subsequently released sediment which became incorporated into the wave.

Author / date	Study location	Sediment source	Morphological impact
Jacobson and Bobitt Gran, 1999	Current River Basin, Missouri, USA	**Exogenous:** Low intensity landscape disturbance resulting from land use change	Low amplitude waves are concentrated into disturbance reaches separated by more stable reaches. Disturbance reaches possess high rates of channel migration of up to 250 m in 50 years (Jacobson and Pugh, 1997). Gravel bar area increased in disturbance reaches with the location of area growth only weakly associated with tributary junctions.
Wohl and Cenderelli, 2000	North Fork Poudre River, Colorado, USA	**Endogenous:** 7000 m^3 from a reservoir sediment release	Deposition of material was primarily restricted to pools. Initial sediment reworking creasted a deep, narrow thalweg. Sediment deposition occurred in lateral eddies. Wave translated as sediment moved from pool to pool as previously reported in Lisle and Hilton (1992) and Wohl et al., (1993).

Table 2.8 Hierarchical bedform classification for uncohesive sediment: After Nicholas *et al.* (1995). Note: t_e is the time taken to pass through a reach (event time).

Bedform Class	Space Scale	Timescale	Typical Features
Microforms	Grain size	$<<t_e$	Single clasts
Mesoforms	Flow depth	$=t_e$	Particle clusters
Macroforms	Channel width	$\geq t_e$	Unit bars
Megaforms	Several channel widths	$>>t_e$	Bar assemblages

Table 2.9 Classification of sediment slugs after Nicholas *et al.* (1995).

Slug Scale	Dominant Controls	Impact on Fluvial System
Macroslug	Fluvial process-form interactions	Minor channel change
Megaslug	Local sediment supply and valley-floor configuration	Major channel change
Superslug	Basin-scale sediment supply	Major valley-floor adjustment

Endogenous and exogenous sediment waves can originate for a range of reasons. These origins can be divided into two main types:

1. natural causes: e.g. landsliding (Beschta, 1983) and channel processes (Wathen and Hoey, 1998)
2. man-induced causes: e.g. mining (Gilbert, 1917; Knighton, 1988; James, 1999), forestry (Madej, 1982; Madej and Ozaki, 1996), distributed land use changes (e.g. Jacobson and Bobbitt Gran, 1999) and sediment flushing from dams (Wohl and Cenderelli, 2000).

In the context of this research, exogenous sediment waves resulting from man-induced sources are of the greatest interest and, in particular, sediment waves resulting from land use change and forestry are clearly of significance. These man-induced waves can further be categorised according to the spatial distribution of the sediment delivery to the channel network producing two broad types of wave which may be expressed very differently in the channel:

1. Point source waves – i.e. landslides, mining debris
2. Non-point source waves – i.e. distributed land use changes, some forestry inputs

2.8 Point source waves – description and morphologic expression

Point source waves originate from elevated sediment supplies which enter the channel over short, discrete reach lengths. Most commonly supply is a result of natural events such as landslides, however, man-induced sediment from mining and sediment flushing from dams are also examples of point source sediments from which such waves have been reported. A simple single dimensional model of point source wave evolution over successive transport events is provided by Nicholas et al. (1995) and presented in figure 2.6. The model is simplistic in that it does not account for downstream waveform changes resulting from interference at tributary junctions and temporary storage (Benda and Dunne, 1997), however it serves as a useful tool for the visualisation of sediment wave evolution. It shows that sediment wave translation is accompanied by wave attenuation, and that these processes are closely related to the frequency of transport events. It also infers that the morphological impact of a point source wave will be most acute, but most short lived, close to the location of the sediment input. Conversely, downstream of the sediment input location the morphological impact of a wave is likely to be less but have a greater residence time within the channel network.

It has been suggested by the flume based studies of Lisle et al. (1997) that not all sediment waves translate downstream according to the model of Nicholas et al. (1995). Lisle et al. found that following sediment input to the flume the wave essentially dispersed both upstream and downstream without translation (figure 2.7).

Figure 2.6 The downstream evolution of a sediment wave as modelled by the single dimensional gamma function model of Nicholas et al. (1995). P(x) is the probability of sediment travelling to a downstream location (x) over a number of transport events (n). Note that the wave both translates and attenuates with increased transport events. From Nicholas et al. (1995).

Figure 2.7 Evolution of a non-translating, flume generated sediment wave (Lisle et al., 1997). The predicted line is that generated by a simple, single dimensional sediment routing model. The wave is seen to disperse symmetrically about the injection point, thereby attenuating without translation. From Lisle et al. (1997).

Sediment that overpassed migrating bars was apparently responsible for dispersal of the wave. Whilst the downstream dispersal of the wave is easily conceived, it is difficult to envisage a mechanism for upstream dispersal in a natural channel raising suspicions that the result may be consequence of studying bed load transport in an artificial flume environment. Indeed, such an upstream dispersal of sediment has yet to be reported in field-based studies. In a situation of a non-translating wave in the channel network one would expect a single reach, at the location of sediment input, to experience morphological adjustment, with the magnitude of that adjustment decreasing with time after the sediment input event.

In field studies of sediment waves the translating wave model is best supported. Gibert (1917) first surmised from the downstream patterns of aggradation and degradation of channel elevation in the Sacramento Valley that the passage of a wave, translating downstream, occurred. More recently the downstream pattern and timing of channel change has also been used to detect translating sediment waves. Madej (1982) shows, via repeat cross profile surveys, a sediment wave generated as a result of timber logging between 1850 and the 1930s. By 1970 the morphological expression of the wave is greatest at the downstream most end of the catchment indicating extensive translation from source areas located upstream. Successive channel surveys between 1970 and 1977 (figure 2.8) show the wave actively undergoing translation. Similar strong evidence of wave translation and attenuation is also provided by Madej and Ozaki (1996). Again repeat cross profiles, surveyed between 1973 and 1991, are used to temporally investigate changes in channel morphology of Redwood Creek, California, due to a forestry-induced wave. Selected sections together with cumulative bed elevation plots are presented in figure 2.9. It is clear, particularly for the cumulative plots, that a sediment wave can be traced moving downstream and that the amplitude of the morphologic change is greatest close to the sediment inputs in the upper reaches and declines progressively downstream.

Figure 2.8 Cross profile change along the Big Beef Creek, USA following an influx of sediment due to logging operations between 1850 and 1930. Note that the uppermost profile experiences little change over the survey period indicating that the sediment has been transported out of the upper reaches and demonstrating wave translation. From Madej (1982).

Figure 2.9 Channel changes in the Redwood Creek showing the translation and attenuation of a sediment wave between 1973 and 1991. Example cross profiles are presented on the left and cumulative elevation change plots are presented on the right. The decrease in wave amplitude with increase in translation is well represented by the cumulative plots. From Madej and Ozaki (1996).

2.9 Non-point source waves – description and cumulative effects

Non-point source waves originate from enhanced sediment supplies which enter the channel over greater reach lengths than point source waves. Jacobson and Bobbit Gran (1999) provide urbanization, agriculture and timber production (all anthropogenic sources) as examples of non-point source inputs capable of generating non-point source waves. In respect of elevated sediment yields from UK upland afforestation, where enhanced bed load yields enter the channel network from the numerous, and spatially distributed confluences between forest drainage ditches and the main channel, any resulting sediment waves can certainly be thought of as non-point source. As individual sediment inputs, forest drainage ditch bed load inputs would be virtually impossible to detect morphologically and would be unlikely to have significant impact on channel morphology over periods greater than the single event. However, non-point source inputs may have a significant impact if they are able to accumulate. Consequently, non-point source waves may be thought of as waves, with the potential to generate a morphological expression and persisting for periods greater than the single event, which originate as a result of within network processes which have acted to cumulatively combine small, distributed sediment inputs to the channel network. They will not necessarily be detected morphologically close to the sediment source areas, but instead will be detected at, and downstream of, the sites of accumulation. These sites of accumulation may be some distance from the sediment source. The processes which control such accumulations are collectively termed 'cumulative effects' (Jacobson and Bobbit Gran, 1999). Understanding the cumulative processes, and their temporal and spatial nature, forms an additional consideration required to fully understanding channel changes resulting from non-point source waves.

2.9.1 Cumulative effects

Distributed sediment inputs may combine due to one of two reasons:

1. As a result of the character of the channel network, distributed sediment supply becomes focussed into a finite reach to the extent that the upstream sediment

supply to the reach exceeds the downstream bed load transport of the river (e.g. Jacobson and Bobbit, 1999). Such a concept implies that bed load transport out of the channel is either restricted in some way at the downstream margin of the affected reach or that sediment supply to the upstream margin of the reach is greatly elevated, for example from tributary inputs. Where the sediment accumulates one might expect bed aggradation, increased gravel bar area and channel widening.

2. Bed load transport rates decline within a finite reach to the extent that upstream sediment inputs to the reach exceed sediment output from the reach.

2.9.1.1 Cumulative effects due to channel network patterns

Church (1983) investigated the processes leading to reaches experiencing high levels of sediment accumulation on the Bella Colla River, Canada. Although his study is not primarily concerned with the cumulative effects of non-point source sediments, it does serve as an important field example of how sediment may accumulate within finite reaches. Church's essential question asked why the downstream channel pattern of the Bella Colla River was not orderly, but one of discrete, laterally unstable reaches with high levels of stored sediment interspersed with stable reaches with low levels of stored gravel (i.e. the 'wandering gravel channel' setting of Neill, 1973). He concluded that

'diffusion [of bed material] must be limited in rivers of Bella Colla type since sediment is reconcentrated in storage areas' (Church, 1983, p. 180)

Church noticed that the loci of sedimentation reaches appeared spatially linked to tributary inputs and, in particular, that alluvial fans which developed at the confluence between the tributaries and the main channel acted to restrict the sediment supply downstream. Consequently, sediment from upstream accumulated behind the tributaries generating sedimentation zones.

Jacobson and Bobbitt Gran (1999) investigated a similar channel pattern type in the Current River, Missouri which was thought to be a result of the cumulative effects of non-point source inputs of bed material resulting from basin land use change. Their study primarily used gravel bar areas, measured from aerial photography, as an index for stored gravel and investigated three hypotheses for the accumumulation of bed material in what they termed 'disturbed reaches':

- Accumulations of sediment may be attributed to increased inputs from tributary basins (i.e. the Church (1983) model).
- Accumulations of sediment may be attributed to valley scale characteristics that would promote longitudinal zones of storage (i.e. meander bend frequencies which may be expected to have a spacing of between 11 and 16 times the channel width (Leopold et al., 1964)).
- Accumulations arise from the channel network controls on the time of arrival of sediment routed from uniform disturbance in the basin.

In contrast to Church (1983), Jacobson and Bobbitt Gran found no spatial correlation between the location of tributary inputs and disturbed reaches. Furthermore, when they investigated the downstream distribution of disturbed reaches they discovered that it did not adhere to conventional concepts of longitudinal variance in valley-scale storage characteristics. Instead, Jacobson and Bobbitt Gran were able to demonstrate that the location of disturbed reaches could be better explained by a sediment routing model which investigated the arrival time of sediment from head water source areas in which it was assumed that the transport time of the sediment from source to disturbed reach was a simple function of the channel distance. Sediment yielded at the same time from headwater erosion located proximal to a disturbed reach would arrive at the reach before that yielded from headwater erosion distal to the disturbed reach. Essentially, their work demonstrates that the channel network pattern, when combined with historically variable patterns of land use change within the basin may, in itself, be sufficient to produce a cumulative effect purely as a result of the timing of sediment arrival to zones of the main channel.

2.9.1.2 Cumulative effects due to decline in bed load transport rates

Bed load transport is of primary importance in understanding the rate and pattern of downstream translation of sediment waves. Bed load transport equations exist which relate sediment transport per unit width to either shear stress, discharge per unit width or stream power per unit width (unit stream power). Bagnold (1980) presents an equation which relates bed load transport to unit stream power (equation 2.1). Data from a range of fluvial environments support this idea and demonstrate a correlation between bed load transport and excess unit stream power (e.g. Hassan and Church, 1992; Knighton, 1998). Gomez and Church (1989) conclude that although none of the ten bed load transport equations they reviewed provided a fully satisfactory prediction of bed load transport rate, equations based on stream power were preferred. As a consequence, understanding of the transport and deposition of sediment in a particular channel may be significantly enhanced with knowledge of the downstream patterns of unit stream power.

$$q_{sb} = (\Omega - \Omega_{cr})^{3/2} d^{-2/3} D^{-1/2}$$

(equation 2.1)

Where

q_{sb} is the sediment transport rate per unit channel width

$(\Omega - \Omega_{cr})$ is excess stream power per unit width (W m^{-1})

d is flow depth (m)

D is grain size (mm)

Lawler (1992) examined downstream change in unit stream power with reference to Welsh river channels. He was able to demonstrate that if a river's downstream increase in discharge can best be described by a power function, and the relationship between distance from source and channel gradient is best described by a negative exponential function, stream power will peak in the middle courses of rivers. The relationships are presented in figure 2.10. In addition, Lewin (1982) concludes that

'generally speaking both stream power and rates of floodplain reworking reach a peak in the middle courses of Welsh rivers: stream power is less upstream where discharges are small and may be less downstream where gradients become minimal' (Lewin, 1982, p. 24).

Figure 2.10 Theoretical mid-channel peak in unit stream power which results when downstream increase in discharge is best fitted by a power function relationship and downstream change in slope is best fitted by an inverse exponential function. From Lawler (1992).

The implication of this for the accumulation of non-point source sediment in sediment waves is great. A peak of unit stream power in the mid course followed by a subsequent decline would, in a situation of unrestricted upstream sediment supply, cause a decrease in the bed load transport rate out of the middle reaches and the storage of sediment in the reaches immediately downstream of the mid course. In a scenario where non-point source sediment is transported downstream in small 'packets' of sediment, slower moving packets immediately below the peak in stream power would be caught by faster moving ones upstream causing their accumulation into a possible sediment wave. Where upstream sediment sources for waves are diffuse (i.e. afforested catchment drainage ditches) such a mechanism provides a feasible model by which non-point source sediments may combine to form macroscale features or larger and hence drive lowland channel change.

2.10 UK upland forestry and sediment wave generation – summary

The sediment wave model is commonly accepted as a means of explaining the downstream movement of elevated sediment yields through a channel network. Sediment waves have been identified in numerous field studies including several studies where elevated sediment yields can be attributed to land use change; particularly catchment forestry. It would therefore seem reasonable to consider it likely that elevated bed load yields following UK upland catchment afforestation may be expressed as sediment waves. However, the non-point source nature of forest inputs would require that cumulative processes be present in the channel network in order for sediment accumulation capable of producing extensive morphological adjustment to occur. If no cumulative processes are present within the channel network it would seem unlikely that elevated bed load yields from UK upland forestry would initiate extensive lowland channel change.

2.11 Lowland channel response to modified peak discharge

The response of a fluvial system to an increase in discharge is shown in figure 2.5. Following an increase in discharge, stream power, a multiplied function of discharge and channel slope, is shown to increase together with channel width and depth. The increased stream power increases the sediment transport rate, which in turn has the ability to affect bed form geometry and channel planform as well as the downstream translation rate of any sediment waves within the fluvial system. The figure serves to demonstrate the complexity of the interrelationships between the responses, many of which can act as feedback mechanisms, both positive and negative. It also serves to demonstrate that river sediment dynamics are heavily influenced by discharge. In the following paragraphs a number of studies of channel responses to increased peak discharges are reviewed, particularly in terms of the impact of increased flow on channel width and depth.

2.11.1 Channel change and increased peak discharge – urbanisation as an analogy

There are two main ways in which discharge can be affected: rainfall / runoff variation as a result of:

1. land use change
2. climatic change

In mid-Wales medium-term land use change has commonly been upland afforestation, the hydrologic consequences of which have already been examined. Catchment afforestation has been shown to have little impact on runoff except during the ground preparation and planting phase when peak discharges have been reported as having increased by up to 40 %. However, there are no studies within the UK literature concerned directly with lowland channel response to the increased peak discharge which has resulted from this phase in the upland forest rotation. However, there is considerable similarity in the hydrograph response of forested catchments in the ground preparation phase and the hydrograph response to urbanisation (figure 2.11). As a consequence, the response of river channels to urbanisation is examined here with the impact of urbanisation on the flood hydrograph being considered analogous to the impact of forestry ground preparation and planting.

Figure 2.11 Unit hydrograph response to (a) forest ditching in the Coalburn catchment, Pennines and (b) to urbanisation along Cannons Brook, Essex. Note that both hydrographs experience a decrease in lag to peak, an increased peak discharge and a reduction in the recessional limb. After Stott (2000).

The process of urbanization is seen to increase peak discharge magnitudes through the spread of impervious areas, which increases the volume of runoff. Lag times are reduced as a response to storm drainage systems which facilitate the fast movement of water through the basin. These responses can be extremely similar to forestry during the clear felling and immature canopy phases of the forest rotation. Knight (1979) was able to demonstrate an increase in peak discharge for Stevenage Brook and the River Tawr of 40% and 18% respectively. These figures are very similar to those presented by Robinson (1980) for the afforested Coalburn Catchment directly after forest ground preparation. The impervious urban areas facilitate runoff in a similar way to the non vegetated ground of a clear felled upland catchment and forest drainage ditches act in a similar way to storm drains concentrating runoff and ensuring fast delivery to the river channel thereby reducing lag time.

It is therefore proposed that examination of lowland channel response to hydrologic change following catchment urbanization will provide a general proxy for the likely lowland responses to hydrologic change following afforestation. There are two major concerns however:

1. That the impact of upper catchment afforestation on the flood hydrograph may be buffered considerably by tributary inputs from non afforested sub catchments by the time the flood wave reaches the lowland channel. Such a buffering does not usually occur in urbanized settings where the responding river reaches are commonly located close to the urbanized area.

2. That enhanced peak discharges from forest preparation and planting will be short-lived, existing only until the forest canopy has gained enough maturity to provide significant interception and evapotranspiration of rainfall. Consequently, the impact on lowland channels of increase peak discharges will probably also be short lived unless flood events of a significant magnitude to induce a new channel equilibrium state occur over the preparation and planting period. Urbanisation, however, permanently alters the flood hydrograph and, therefore, has a greater potential to cause medium or long-term channel responses.

2.11.2 Response of width and depth to urbanization

Hammer (1972) studied the impact of suburban development in 78 watersheds in Pennsylvania on peak discharges and the subsequent responses of Piedmont channels. Where urbanization increased peak discharge, channel capacity was reported to increase through a mixture of channel widening and degradation, although no figures are offered. Perhaps more importantly, Hammer was also able to demonstrate that the increase in channel capacity was positively related to the slope of the river channel and negatively related to the distance between the source of increased peak discharge and the responding channel. The latter point is of particular importance for this study where the source of increased peak discharge is a considerable distance upstream of the lowland reaches and infers that by the time forest runoff has reached the lowland channel its impact on the lowland flood hydrograph may be reduced.

In the previously mentioned study of Knight (1979), Stevenage Brook increased capacity mainly via the depth dimension whilst bed rock restriction in the River Tawd forced capacity to increase mainly via the width dimension. The impact of bedrock on the River Tawd is important suggesting that the mechanical strength of channel boundary material may act to control the dimension within which the channel may respond. Park (1980) also investigated the responses of width and depth to increased peak discharge following urbanisation of the River Deer and Woodbury Stream in Devon. The study compares measured channel width and depth to that predicted by power function relationships which describe the relationship between drainage basin area and channel width and depth. Neither channel is bed rock restricted in the depth dimension and consequently the depth enlargement ratio was greater than that for width in both catchments. The River Deer exhibited a depth enlargement ratio of 135% and a width enlargement ration of 106% whereas Woodbury Stream responded to urbanisation with a depth enlargement ratio of 182% and a width enlargement ratio of 141%.

The results from the above studies provide evidence that an increase in lowland peak discharge can be associated with an increase in lowland channel capacity, via adjustment of channel width and depth. However, in the case of the rivers where the source of increased discharge is distal to the responding reach buffering of the flood wave may occur and responses may be considerably reduced.

2.11.3 Other studies

Kellerhals et al. (1979) examined the morphological responses of Canadian river channels in which discharge had been altered in response to inter-basin river diversions. Of the 19 diversions studied, 4 have documented evidence of increased peak discharges and associated morphologic change. These are summarised in table 2.10.

Perhaps the most important feature of Kellerhals et al's results is the difference in response between river channels whose discharge has been elevated for several decades and those for which the altered discharge has been a recent feature. In both rivers where elevated discharge has been recent no channel change was observed, despite a five-fold increase in the two year flood at Poplar Creek. At Cheslatta River, however, where the two year flood discharge had increased 11 times, an entirely new river channel had been formed, perhaps representing the establishment of a new stable equilibrium channel. These results are significant in terms of lowland channel changes in response to short-lived elevated peak discharges following planting and ground preparation of upland forestry. They suggest that the impact of a 40 % rise in peak discharge over a period of only four or five years is likely to be small.

Table 2.10 The impact of inter-flow diversion of discharge and channel morphology.

Q_{mean} = mean annual flow (cumecs)

$Q_{2\,year}$ = 2 year flood (cumecs)

Q_{max} = highest observed flow (cumecs)

Stream name and period of altered flow	**Natural flow**		**Diverted flow**		**Channel response**
Cheslatta River	Q_{mean}		Q_{mean}	138	New gravel bed river channel
25 years	$Q_{2\,year}$	3.0	$Q_{2\,year}$	331	incised 5-10 m below
	Q_{max}	?	Q_{max}	507	floodplain
Poplar Creek	Q_{mean}	0.5	Q_{mean}	2	
3 years	$Q_{2\,year}$	4.5	$Q_{2\,year}$	24	None observed
	Q_{max}	5.5	Q_{max}	54	
Rat River	Q_{mean}	118	Q_{mean}	883	
3 years	$Q_{2\,year}$	330	$Q_{2\,year}$	?	None observed
	Q_{max}	578	Q_{max}	963	
Little Jackfish River	Q_{mean}	5	Q_{mean}	117	Persistent degradation and
36 years	$Q_{2\,year}$	?	$Q_{2\,year}$	320	bank erosion
	Q_{max}	65	Q_{max}	450	

In the UK, a study by Knighton (1973) on the Bollin-Dean, a gravel-bed river in Cheshire, demonstrates that the flood peak pattern is also an important factor which influences channel change, particularly bank erosion. He states that:

'Summer storms of high intensity but short duration generated single peak flows which, unless of a large magnitude, caused little or no bank erosion. In contrast, winter storms produced more complex multi-peaked flows which, although they were individually of lower magnitude, were more effective in eroding the channel banks.' (Knighton, 1973, p. 402)

Unfortunately, the impact of forestry on flood peak complexity remains poorly understood, however, it would seem unlikely that it would have a large impact, particularly as increased flood peak complexity in winter is linked to the higher duration of storm events rather than catchment characteristics. However, the study does demonstrate that the potential for elevated peak discharges to cause channel change is likely to be higher in the winter months.

2.11.4 Discussion

The above studies serve to outline the general pattern of response of width and depth to an increase in peak discharges. These responses are, however, commonly based on studies where the discharge has been altered for a period in excess of 10 years. Robinson (1980) identifies enhanced peak discharge as a result of afforestation during the clear felling and immature canopy phases only, representing a period of perhaps 3-4 years in every 50. For the remainder of the rotation peak discharges remain unaffected or even attenuated. This is considerably different to the response of bed load yields which remain elevated throughout the forest rotation. Such a pattern of response provokes questions about timescales of adjustment for width and depth.

Knighton (1998) schematically represents timescales of adjustments of the major fluvial variables (figure 2.12). It is clear from this diagram that planform change operates over a timescale in excess of 10 years. As a result it is extremely unlikely that the short term increase in peak discharge as a result of forestry would influence

Figure 2.12 Schematic diagram representing the response times of fluvial parameters of varying scale. Note that over the medium-term (5-50 years) channel width, depth and bed configuration could be expected to undergo change. From Knighton (1998).

lowland planform. Hence sediment loads are far more likely to be of importance. It is, however, possible for channel width and depth to adjust over far shorter timescales, even over a single major flood event. Consequently, enhanced peak discharge may promote some cross sectional adjustment.

2.12 Medium-term lowland channel response to upland afforestation – summary

Of clear importance regarding the temporal pattern of adjustment of a lowland fluvial channel is the temporal pattern and duration of change in the sediment loads and runoff regime of the channel network. A short-term change, followed by a return to the original conditions, will provoke only a short lived channel response from morphologic variables able to adjust quickly - namely the channel cross section and bed configuration. The identification of such changes will require surveys of high temporal and spatial resolution. Longer-term, sustained change may be identified from adjustments of planform and channel gradient.

There is little doubt that upland afforestation has the potential to provide significant sustained medium-term enhancement to lowland bed load yields and that these loads will enter the channel network as non-point source inputs. The pattern of downstream variation in unit stream power in Welsh rivers provides a possible mechanism for accumulation of bed load in the middle reaches of channel networks, and hence the development of non-point source waves. Consequently, it may be possible for lowland channels, downstream of forested upper catchments, to display zones of instability analogous to Jacobson and Bobbitt Gran's (1999) 'disturbed reaches' caused by non-point source sediment waves. In such reaches one might expect significant morphological adjustment including increased gravel bar area, bed aggradation, and lateral instability. Lowland channel adjustment occurring in response to forest derived bed load will be a lagged response, the lag time being determined by the time taken for forest-derived bed load to be transported through the channel network to the site of sediment accumulation.

The consequences of short periods of enhanced peak discharge are difficult to predict. In the absence of a triggered feedback mechanism it is unlikely that 3-4 years of enhanced peak discharges, even of 40%, will alter channel geometry significantly over the medium-term, although there may be localized short-term effects. Where the channel does respond, an increase in both width and, more importantly, depth may be recorded. Planform would be unlikely to adjust. Unlike the response of channel geometry to sediment yields, in which there is a lag time between sediment release and lowland response, lowland adjustment to hydrologic change may be rapid.

It should also be noted that the role of modified hydrologic inputs to the fluvial system as a result of medium-term increases in rainfall may be important in driving lowland channel change. As a consequence rates of lowland channel responses should be considered in the context of medium-term hydrologic records. Where both bed load and hydrologic inputs to the channel vary over the same period as channel changes occur it is potentially extremely difficult to distinguish the relative influence of each.

2.13 Statement of the problem and study site selection.

This project is concerned with the impact of UK upland catchment afforestation on the lowland channels of the rivers draining them. As previously shown, there is a considerable body of work outlining upland channel responses to forestry, however there is no such body of work for lowland channels. This is surprising given the greater economic value of lowland floodplains and river channels. This lack of literature is, in part at least, due to the fact that the transport of elevated bed load yields to the lowland channels, several kilometers downstream of the upland forest, may take in excess of several decades. With the majority of upland afforestation having occurred in Scotland between 1970-1974, England between 1955-1959 and Wales between 1960-1964 (Omerod and Edwards, 1985) it is only now that lowland channels may be expected to respond.

The studies cited in this chapter have provided evidence that lowland channels may indeed undergo substantial morphologic adjustment in response to upland catchment bed load yield increases, particularly if cumulative effects are acting on the lowland channel network to promote the accumulation of non-point source sediment waves. In response to such waves, lowland channels have been reported as having increased width, a higher gravel bar area and being laterally unstable. Leeks et al. (1988) describe patterns of instability on the lowland reaches of the Afon Trannon in mid-Wales which are similar to the patterns one might expect following the accumulation of non-point source sediment. Using studies of historical data sources, Leeks et al. (1988) show that, at all four field sites studied, mean and maximum bank erosion rates increased between 1901-1948 and 1948-1976. At one site bank erosion rates are recorded as increasing from a historical rate of 0.3 m yr^{-1} to a rate between 1982 and 1983 of 0.71 m yr^{-1} and at three of the four sites maximum erosion rates had increased continuously since 1885. In concluding the reasons for this increased lowland instability of the river Leeks et al. state that,

'the Trannon's wide, deep glacial trough and the relatively recent afforestation of the headwaters are longer and shorter-term aspects of relevance' (Leeks et al., 1988, p. 222)

No direct evidence of the importance of the upland catchment afforestation is provided in the study, however circumstantial evidence suggests that upland forestry would indeed appear to have a role in the recent destabilization of the Trannon. The afforestation of the upland catchment began in 1948 – a date which coincides with the elevation of bank erosion rates in the lowland channel. Moreover, Leeks et al. (1988) report that sections from the Trannon, surveyed by Severn Trent Water Authority in 1985, show the build up of bar forms in the lower reaches indicative of increased bed load supply. Their own surveys also confirm this with a shoal in the lowland reach reported to have increased in volume by 2.55 m^3, deflecting flow and generating 20.94 m^3 of bank erosion.

This study aims to investigate, in further detail, the medium-term patterns of channel change in the lowland Trannon, and to attempt to determine the extent to which the magnitude, temporal and spatial patterns of medium-term lowland channel changes can be explained by sediment loads and streamflow changes as a result of upland catchment afforestation. It aims to discuss the implications of the findings both in terms of the Afon Trannon, and more broadly in order to conclude the extent to which changes in upland catchment land use can influence lowland channel stability.

3. AIMS AND OBJECTIVES AND STUDY SITE DESCRIPTION

3.1 Aims and objectives

The overriding aim of this study is to investigate the extent to which the afforestation of the upland catchment of the Afon Trannon can be held responsible for medium-term patterns of lowland channel instability, and the major geomorphological processes operating to cause this instability. As such, the study has necessarily had to:

- Characterise and quantify the historical patterns of channel change of the Afon Trannon both in the medium-term (5-50 years) and longer-term (> 50 years) time frames.

- Characterise and quantify the short-term (< 5 years), contemporary patterns of channel change of the Afon Trannon and where possible identify and investigate the processes driving these changes.

- Examine the channel changes and channel processes in respect of the medium and short-term streamflow charcteristics and sediment (in particular bed material) loads of the river.

To achieve this, the following four objectives will be addressed:

1. The generation of a medium-term bed load yield estimate for the forested upper catchment.

From the preceding chapter it is apparent that forested UK catchments have elevated bed load yields and that this bed load, if yielded in sufficient quantity, has the potential to drive channel change. Consequently, if the importance of forest derived bed load as an agent for the lowland channel changes of the Afon Trannon is to be established, quantification of medium-term patterns and magnitudes of the bed load yield of the upper Trannon catchment is essential.

Whilst there are considerable physiographic data and forest operations data available for the upper Trannon catchment, no data pertaining to the impact of catchment forestry on bed load yields exists and no quantification of actual bed load yields has been undertaken. This is in contrast to the neighbouring rivers in the Institute of Hydrology's (currently part of the Center for Ecology and Hydrology) experimental catchments at Plynlimon. Here physiographic and forest operations data are complemented by bed load yield records which extend over three decades. In order to generate a medium-term bed load yield estimate for the upper Trannon, extrapolation of bed load yields from the Institute of Hydrology experimental catchments to the upper Trannon catchment are undertaken based on catchment physiographic comparison and the use of empirical formulae which relate sediment delivery to catchment physiography.

2. The generation of a medium-term hydrologic record for the Afon Trannon.

River channels also respond to changes in streamflow patterns. As such, the investigation of medium-term river channel changes must be placed in a hydrologic context where changes in the medium-term magnitude and frequency of competent flows are quantified. Therefore, the medium-term bed load yield must be accompanied by a medium-term hydrologic record.

The Afon Trannon is poorly gauged. A single stage recorder exists at Caersws, close to the confluence with the River Severn. Here a 15 minute stage record covering the period 1994 – present is available. Prior to 1994 neither stage nor discharge data are available for the channel. Longer-term records of rainfall in the mid-Wales region do exist. In particular, a daily total rainfall record for Dollyd covers the period 1969 – present and this forms the longest rainfall data set recorded proximal to the Trannon catchment. To extend the Caersws stage record rainfall /

runoff modelling will be undertaken using a data-based mechanistic transfer function model (see Young and Beven, 1994). The model input is provided by the Dollyd rain gauge daily rainfall totals and the model output provides a daily mean stage (DMS) record covering 1969 – present. Medium-term changes in runoff regime can then be quantified and the influence of streamflow on historical and contemporary channel changes can be investigated.

3. Quantification of the historical channel changes of the Afon Trannon.

Examining the medium-term history of the lowland channel is vital to this work. Patterns of channel change (e.g. channel widening, lateral channel movement and gravel bar growth) can be used to make inferences about the sediment and hydrologic loads of a river which can then be compared to the estimated loads. Attempts can then be made to correlate the temporal pattern of changes in the river loads with the pattern of land use change within the catchment. Subsequently, tentative conclusions about land use change and historical channel response can be drawn.

Leeks *et al.* (1988) summarise the channel change history of the lowland Trannon for three periods between 1885 and 1976, but their work considers only four short reaches and no estimate of the errors involved in their quantification is offered. Channel change between 1885 and 1948 is measured from historical maps: a method which has a high potential for error (Hooke, 1982). Consequently, the medium-term history of channel change at the Trannon remains relatively poorly quantified.

The recent development of digital photogrammetry, rectification software and GIS has facilitated both quick and accurate measurements of channel parameters from archive aerial photography. These techniques are used to examine a range of channel parameters in both the upland and lowland reaches of the Trannon. The photography to be assessed includes imagery from 1948, allowing channel parameters to be quantified at the point of upper catchment afforestation. In contrast to the study by Leeks *et al.* (1988), this study will examine extensive reaches of the Trannon with a high downstream resolution of measurement at an approximately 10 year temporal interval. This provides a highly detailed record of the temporal and spatial patterns and magnitudes of the channel changes of the Afon Trannon since upland catchment afforestation.

4. Generating a contemporary channel change record and examining the processes driving lowland channel change.

Detailed contemporary field study of channel morphology is undertaken with particular attention paid to lowland reach scale sediment budgets. The budget can be defined as:

$$I = \Delta S + O \qquad \text{(equation 3.1)}$$

where
I = total bed load inputs to the lowland reach (i.e. medium-term upstream and contemporary local)
ΔS = change in the volume of bed load storage
O = total bed load transported from the reach

When constructed, contemporary sediment budgets can be used to assess the relative importance of medium-term upstream sediment inputs (specifically the estimated upland catchment forest inputs) and short-term lowland inputs (i.e. local inputs from bank erosion). The respective influences of each temporal-scale input on changes in contemporary bed load storage and hence the resultant contemporary channel changes may be determined. In this way the influence of medium-term and short-term parameters are evaluated in terms of the contemporary lowland channel system.

This work focuses mainly on the upper-most sites of within-channel bed load storage. Where possible, these sites coincide with reaches for which historical channel change assessment has been undertaken (see above). Repeat cross-profile surveys and a GIS are used to construct accurate, three dimensional models of the channel for which temporal changes in the sediment budget of the reach can be quantified and the mechanisms of channel change can be characterised. Temporal and spatial patterns of sediment transport and deposition are then examined along with discharge records allowing conclusions about the influence of bed load transport processes on channel instability to be made. Finally, investigation of bank material allows the bed load supplied from lowland channel banks to be estimated. Using these data, comparisons can be made between the bed load supplied to the lowland reaches from bank erosion and the bed load yield from the upper catchment.

3.2 The Afon Trannon - overview

The Afon Trannon, its upper catchment, upland forestry and main tributaries are shown in figure 3.1. The river originates in the Cambrian Uplands, Powys, mid-Wales some 15 km east of Machynlleth. The 18.5 km long river has a catchment area of 72 km^2 and is a major tributary of the upper River Severn. The confluence of the two rivers and the Afon Carno occurs close to Caersws (see figure 3.1). The river has been gauged 1.5 km above the confluence at the Caersws stage recorder since 1994. The Trannon catchment is proximal to the Institute of Hydrology's (CEH's) experimental Plynlimon catchments in which the Dollyd rain gauge and Tanllwyth experimental catchment are situated.

The Trannon channel can be divided into several distinct sections (after Leeks *et al.*, 1988):

1. 0-5 km: The river crosses open moorland and a coniferous forest plantation, begun in the late 1940's, known locally as Dolgau Forest. The upper catchment

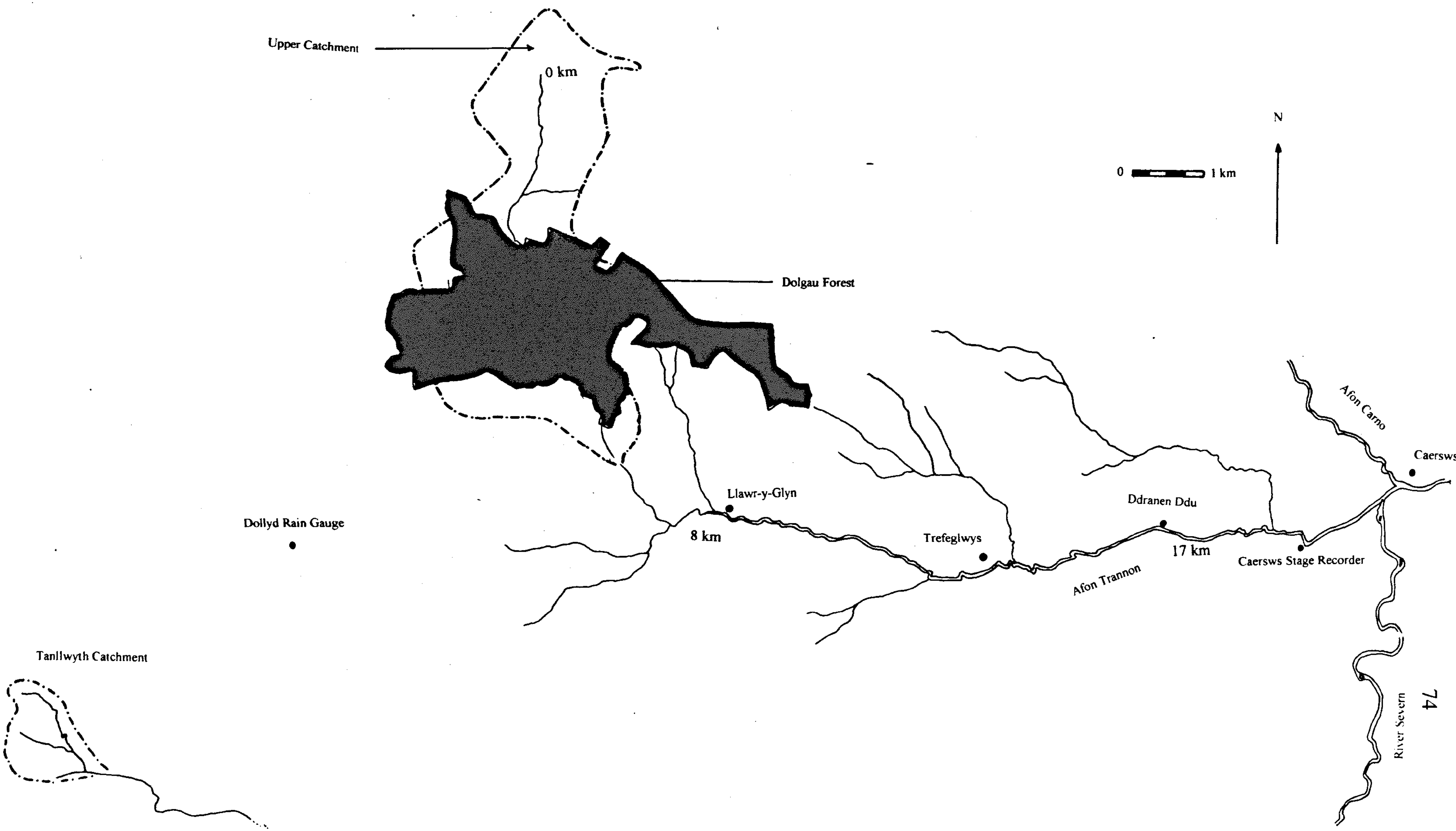
is large relative to the basin as a whole (15 %) and Dolgau Forest extends over approximately 56 % of the upper catchment. Channel gradient is relatively low in the upper most reaches (approximately 0.03). The upper catchment is not gauged.

2. 5-8 km: The river enters a deeply incised, straight, bed rock controlled reach where channel gradient increases sharply to a maximum of 0.051. Bed rock control is provided by friable and highly cleaved Ordovician and Silurian mudstones and shales. The channel is characterised by a high number of waterfalls and large, glacially derived boulders and cobbles with virtually no sediment storage evident.

3. 8-12.5 km: The river enters a slightly more open valley with floodplain width ranging between 150 m and 375 m, and flows eastwards. Gradient ranges between 0.02 and 0.009 and the channel sinuosity remains low at 1.1. The river flows through both bed rock controlled and alluvial reaches. In the alluvial reaches banks are composed largely of sand and silt with occasional sections where composite banks of less than 1 m height are visible. There are few storage sites for sediments and it is thought that bed load transport is rapid through this reach (Leeks et al., 1988).

4. 12.5–17 km: From Trefeglwys downstream the river flows through a broad trough. Gradient ranges between 0.007 and 0.003. The channel is characterised by highly unstable, meandering reaches with composite banks generally in excess of 1 m height, interspersed by straighter, more stable reaches with sand and silt banks. Bed load storage is evident within the meandering reaches in the form of large gravel point bars. Commonly barforms are accompanied by evidence of bank erosion on opposite banks. Borehole information obtained from cores extracted in the downstream end of the trough, and reported in Leeks *et al.* (1988) is shown in figure 3.2. The cores show the valley to be a glacially

Figure 3.1 The Afon Trannon and upper catchment boundary showing the location of study reaches, the Tanllwyth experimental catchment at Plynlimon, Dollyd rain gauge and the Caersws stage logger.



scoured trough infilled with a mixture of till and glacio-fluvial derived gravels and sands. The maximum infill depth reported is 68.6 m.

5. 17-18.5 km: From a point 1.5 km upstream of the confluence with the Severn the channel has been artificially straightened. Bank material is predominantly sand and silt. Bed gradients range from 0.001 to zero. The river stage is monitored at the Caersws stage logger, 1.5 km upstream of the confluence with the Severn.

3.3 Upper catchment background and history

The 10.6 km^2 upper catchment encompasses the upstream most 8 km of the Trannon channel. The majority of the catchment has low topographic relief being dominated by the low rolling hills of the Cambrian Upland plateau. However, at the downstream margin of the catchment the plateau is dissected by a steeply incised valley occupied by the Trannon. The upper catchment has an altitudinal range of between 488 m AOD and 253 m AOD. However, 90 % of the catchment lies on the plateau above 400 m AOD.

The parent material of the upper catchment, and consequently the majority of the Trannon bed load, is extremely similar to that of the nearby experimental Plynlimon catchments being composed of Palaeozoic grits, mudstones and shale or glacially derived drift (Kirby et al., 1991). Stott and Sawyer (1998) analysed abrasion rates of the least resistant shale component of the bed load within the Afon Cyff and Afon Tanllwyth and report mean weight losses of 2.83 % and 4.15 % per year respectively. Therefore, the potential exists for a moderate change in the proportion of bed load and suspended sediment in the Trannon with transport of shale clasts from the upper catchment to the lowland reaches. The abrasion of the more resistant bed load components, of which the total proportion in the bed is not given, is unknown but likely to be far higher than that of the shale fraction meaning that

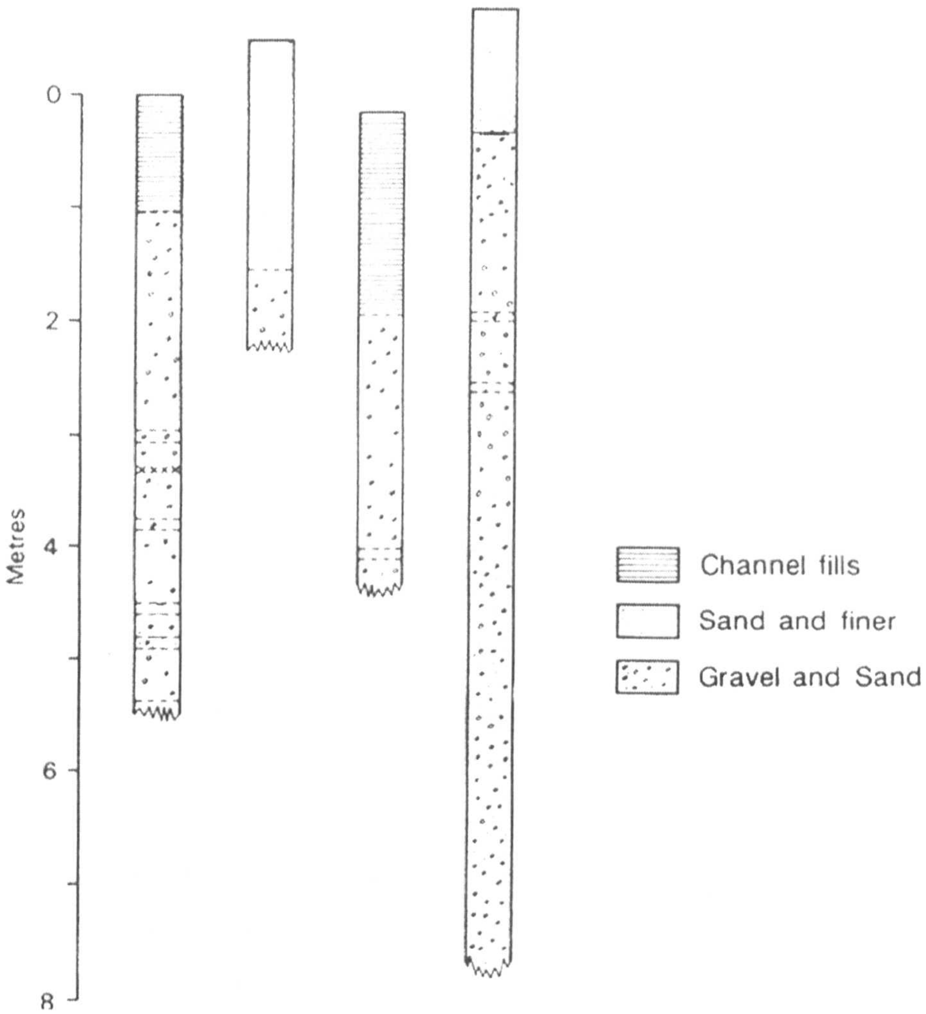


Figure 3.2 Example cores from the lowland Trannon floodplain above the Severn confluence. After Leeks et al. (1988).

overall bed load abrasion rates are likely to be lower than the figures of Stott and Sawyer (1998).

Soil differentiation is mainly dependent on drainage. On the plateau, drainage is impeded and the accumulation of organic deposits, particularly blanket peat of the Crowdy association which have a depth greater than 80 cm, are developed. On more freely draining slopes of the plateau margins soils are podzolized (Rudeforth, 1967) and can be classified as being of the Hafren association. There is evidence of hagging (a form of peat erosion) of the Crowdy association in the Cambrian Uplands (e.g Rudeforth et al., 1984), however there are no reports of peat erosion occurring to the extent that the underlying drift is exposed.

Prior to 1948 the upper catchment of the Trannon was entirely used for rough pasture. Since 1948, however, a major land use change has occurred in the development of Dolgau forest. The forest development has been responsible for land use change in more than 50 % of the catchment. Dolgau forest was developed in two main phases. The planting is summarised in figure 3.3. Between 1948 and 1953 181.6 ha of the upland catchment was drained and planted with Sitka Spruce and Norway Spruce. Between 1966 and 1978 a further 410.9 ha were drained and planted. Many of the earliest forest parcels are now on their second rotation accounting for further planting activity between 1979 and 1993. Herring bone pattern, open drainage ditches, similar to those employed in the neighbouring Plynlimon experimental catchments (see Newson, 1980) were dug during afforestation and are maintained in the forest. These ditches commonly exceed the depth of the blanket peat and have exposed the underlying drifts.

3.4 Lowland Trannon – description and history

The lowland reaches of the Trannon, between 8 km and 18.5 km, flow through improved pasture. Examination of aerial photographs between 1948 and 1995, particularly the 1976 photography of the catchment, shows that some localised

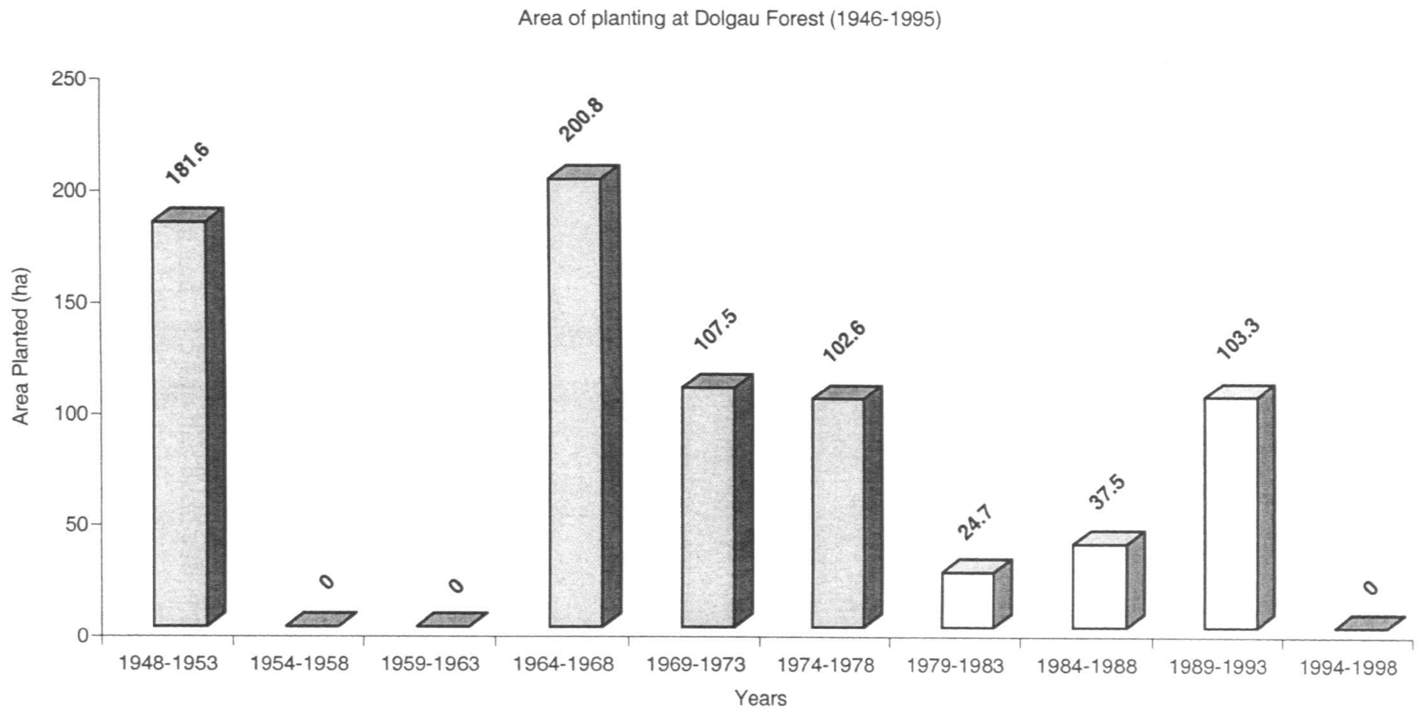


Figure 3.3 Planted areas at Dolgau Forest 1948-1997. Second rotation planting is shown as white bars.

drainage improvement has occurred in the latter half of the 20^{th} century (figure 3.4). Examination of the same photographs offers evidence of more extensive drainage in the catchment prior to 1948. The photograph shows evidence of numerous palaeochannels in the valley floor between 17 and 18.5 km from the Trannon source which Leeks et al. (1988) interpret as evidence for a previously anastomosing channel. The change from an anastomosing to meandering channel has been associated with a change from a wetland and marsh environment to a drier environment (Smith, 1983) and it would seem likely that much of the drainage required to produce this change is a result of man induced drainage associated with pasture improvement. However, it is the author's view that the channel pattern observed in figure 3.4 could also be interpreted as evidence of avulsion.

The use of the lowland Trannon floodplain for cattle grazing is extensive, producing a potential for bank poaching and local instability as a result of the action of cattle. However, Mount et al. (2000) find only a single location at Trefeglwys where bank erosion can be attributed to the action of cattle. Little evidence of poaching by cattle is evident in the field with stock seldom crossing the argaes to reach the channel. Consequently it is the author's opinion that the role of cattle in the lowland stability of the channel is not significant.

In addition to drainage, the lowland channel has a history of channel engineering. Leeks et al. (1988) examined historical maps of the Trannon and report straightening of the Trannon channel, undertaken between 1832 and 1847, between 0 and 4 km from the confluence with the Severn. It is, however, difficult to determine the extent to which this might be a result of channel pattern generalization in the maps, a procedure commonly employed in early cartography. More recently, the channel between 12.5 km and 18.5 km has been subject to extensive engineering works carried out by Severn Trent Water Authority between 1978 and 1979 to alleviate flooding. Planform was left largely unaltered, however the channel section was modified significantly. The main morphological adjustments of the channel were:

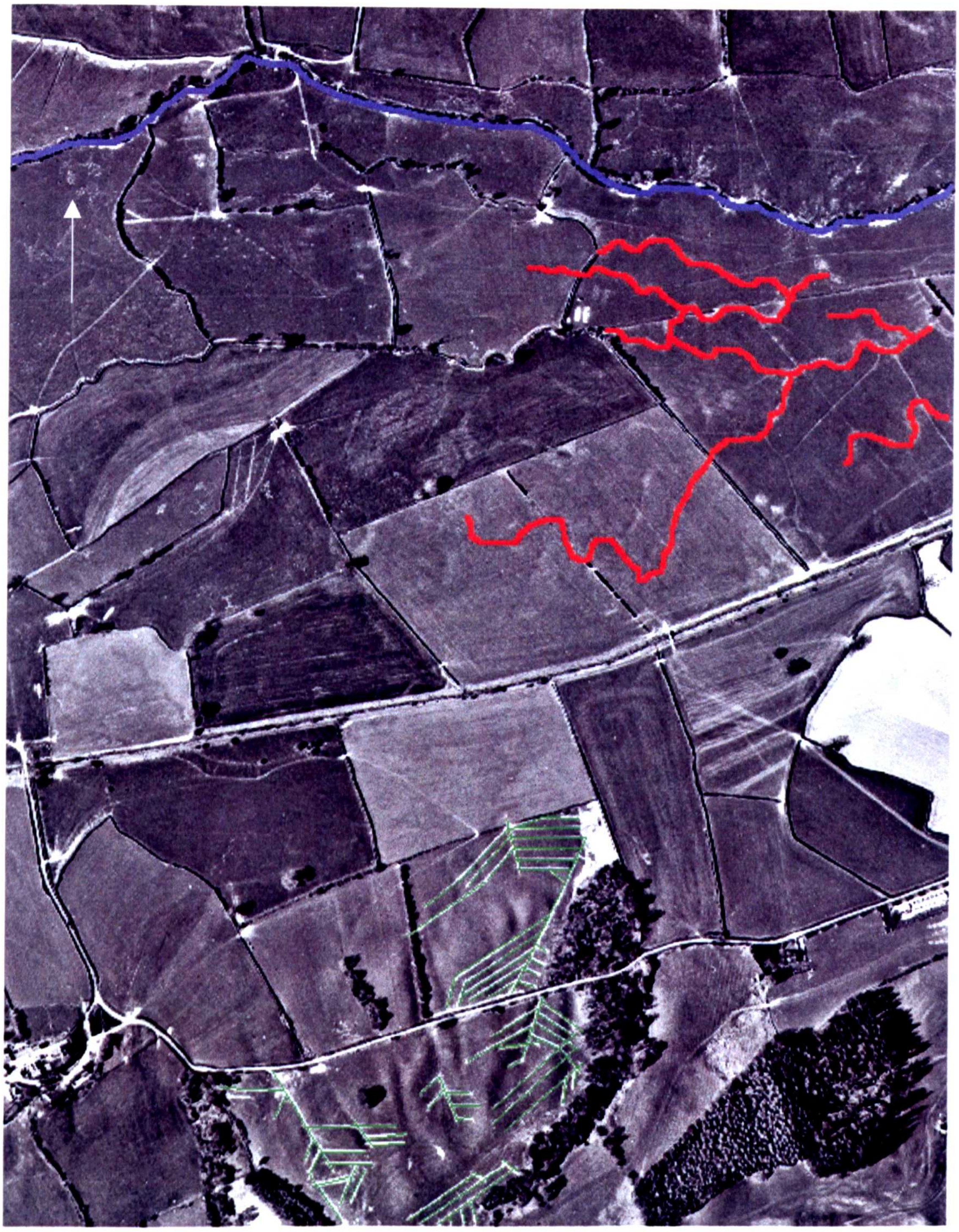


Figure 3.4 Locations of the current (blue) and palaeochannels (red) of the lowland Trannon approximately 2.5 km upstream of the Severn confluence. Recently installed drainage pipes are coloured green. Image scale is approx 1:7600. Arrow indicates north and flow is left to right.

- Dredging of bed forms and grading to a flat floor in cross section
- Establishment of a trapezoidal section with sloping banks achieved by pushing dredgings against naturally occurring vertical bank faces
- Heightening of banks using dredged river material to create argaes - a traditional form of flood bank set back from the channel
- Positioning of gabions on the outside of meander bends
- Positioning of boulders on the outside of meander bends

These engineering solutions were focussed on particularly active reaches and aimed to increase the channel flood capacity by decreasing the flow resistance caused by gravel bed forms and by increasing the effective bank height. In this way stream power during peak discharges was increased. However, the use of poorly cohesive gravels to construct new channel sections was largely unsuccessful, with the increased stream power resulting in increased bed load transport and the erosion of the trapeziodal section to the former channel section. Indeed, Leeks et al. (1988) report extensive failure of the engineering during the winter floods of 1979 – 1980, just one year after completion. Consequently, the remaining engineering in the lowland channel is restricted to gabions and boulders at the outer banks of meanders and the argaes.

3.5 Study reaches

Four study reaches; one at Llawr-y-Glyn, two at Trefeglwys and one at Ddranen Ddu, have been selected for detailed study. The reason for their selection is outlined together with the main characteristics of each reach in table 3.1. The relative locations of the field study reaches at Trefeglwys and Ddranen Ddu are shown in figure 3.5 and described in further detail subsequently.

3.5.1 Chapel View (CV)

The 130 m reach is approximately 13 km from the Trannon source has a bed gradient of 0.005 and a sinuosity of 1.54. The reach exists at the upstream end of the broad glacial trough occupied by the lowland Trannon and represents the upstream most site where large scale bed load storage exists and this storage appears to be associated with rapid bank erosion and lateral instability. The channel contains two active meanders with associated gravel point bars (figure 3.6 and 3.7). Much of the length of the outer banks of the meanders are protected by boulders and gabions from the 1979 engineering work, however two site visits in September and October 1997 confirmed substantial bank erosion along unprotected bank faces. Banks are composite (figure 3.8) and contain 40 % coarse to fine gravel and 60 % matrix of medium sands and clays. D_{50} of bank gravels (material > 2 mm) is 23 mm and the D_{max} is estimated at 70 mm. Where banks are exposed, bank faces are commonly vertical and range between 0.8 m and 1.5 m high. Good aerial photographic coverage exists for this reach.

3.5.2 Bodiach Hall (BH)

Bodaich Hall represents the longest study reach investigated. It is shown in figures 3.9 and 3.10. The reach is 360 m long and located approximately 500 m downstream of Chapel View, approximately 13.5 km from the Trannon source. The reach has a bed gradient of 0.003 and a sinuosity of 1.49. Investigation of oblique photographic records from 1980-1983 and archive aerial photography taken between 1948 and 1995 demonstrated the reach to be laterally unstable (figure 3.11) and to have undergone channel widening, particularly in the period post 1963. Bed load storage is currently evident in two large point bars which form the next location for gravel storage downstream of the CV reach. Bank material varies through the reach from 0 % gravel to 50 % gravel although the majority of the banks contain some gravel. Where composite banks exist the D_{50} of the gravel fraction (> 2 mm) is

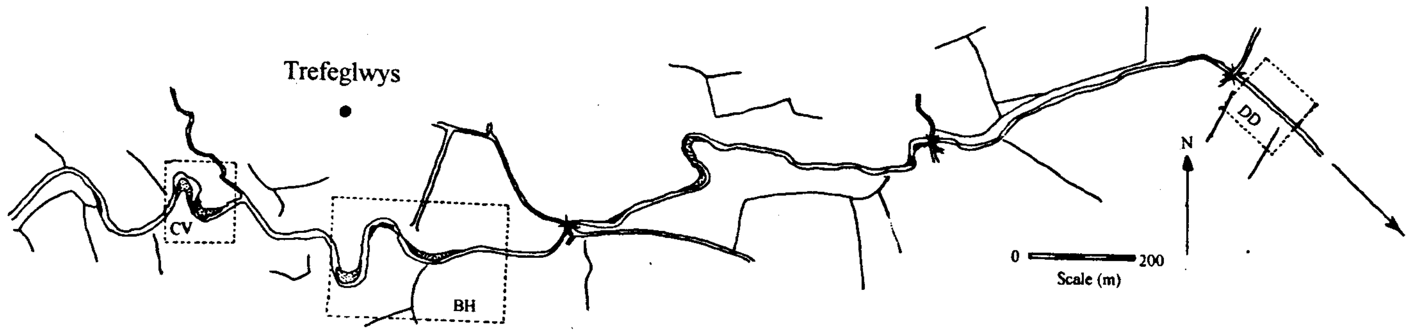


Figure 3.5 The lowland Afon Trannon showing the relative locations of Chapel View (CV), Bodiach Hall (BH) and Ddranen Ddu (DD).

Table 3.1 Major characteristics and reasons for selection of study reaches.

Study reach name and location	Reach characteristics
Llawr-y-Glyn	• 680 m long • Sinuosity = 1.1 • Gradient = 0.02 • Bank material = mainly medium sand and silts with occasional gravel • Riparian vegetation = pasture • Aerial photo coverage = 1948, 1976, 1995
Chapel View (CV), Trefeglwys	• 130 m long • Sinuosity = 1.54 • Gradient = 0.005 • Bank material = gravel and composite material. Gravel fraction: D_{50} = 23 mm, D_{max} = 70 mm • Riparian vegetation = pasture / mature trees • Aerial photo coverage = 1948, 1963, 1976, 1988, 1995
Bodiach Hall (BH), Trefeglwys	• 360 m long • Sinuosity = 1.49 • Gradient = 0.003 • Bank material = mainly gravel and composite material although some finer. Gravel fraction: D_{50} = 25 mm, D_{max} = 70 mm • Riparian vegetation = pasture / low shrubs • Aerial photo coverage = 1948, 1963, 1976, 1988, 1995
Ddranen Ddu (DD)	• 50 m long • Sinuosity = 1.0 • Gradient = 0.002 • Bank material = fine sands / silts • Riparian vegetation = pasture • Aerial photo coverage = 1976, 1988

25 mm. Bank heights range from 0.5 m to 2.0 m, with the majority of bank faces being greater than 1 m high, and cantilever failure is evident.

Channel engineering in this reach is evident from the gabions and boulders which exist along much of the outer banks. However, in the upstream most portion of the reach this bank protection has become incorporated into the inside bank.

3.5.3 Ddranen Ddu (DD)

The Ddranen Ddu reach encompasses approximately 50 m of channel 15.1 km downstream of the Trannon source. It is shown in figures 3.12 and 3.13. The reach is part of a straight, laterally stable section of the lowland Trannon which extends approximately 800 m from the study reach in both upstream and downstream directions. Bed gradient is 0.002. Throughout this region there is little bed load storage visible. Banks are composed of fine sands, silts and clays. No gravel is present in the bank faces. Where bank failure has occurred rotational slumping is the major mechanism (figure 3.13). No recent channel engineering is evident. Poor photographic coverage of the reach prevents photogrammetric analysis.

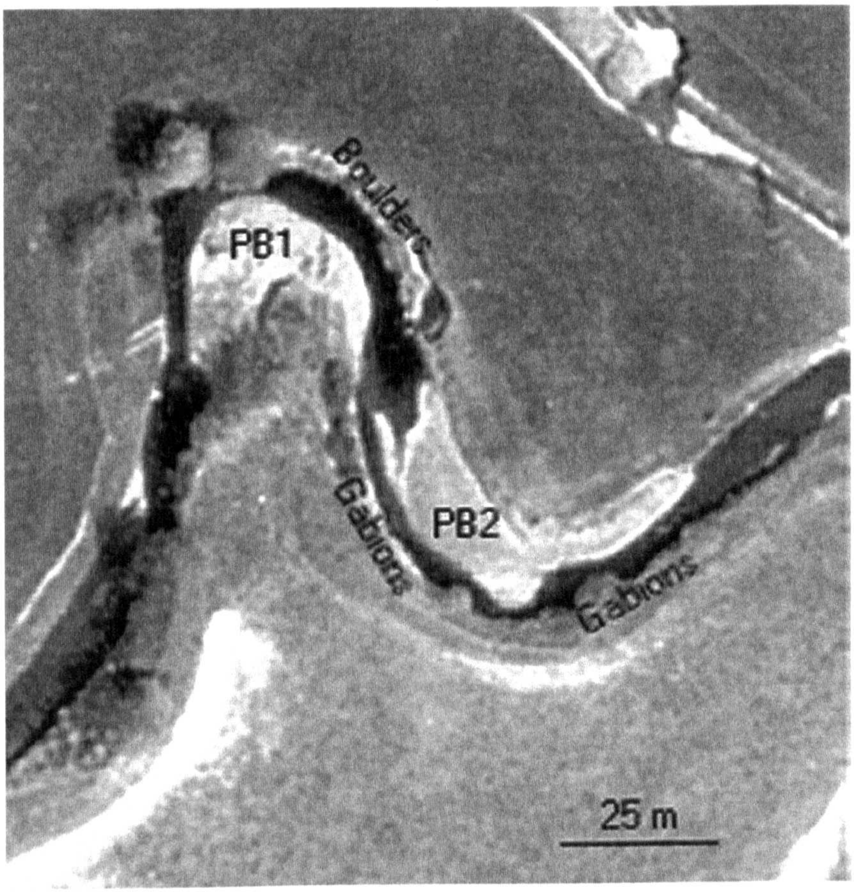


Figure 3.6 Aerial photograph (1995) of the Chapel View study reach showing the location of point bars PB1 and PB2. Also shown are the location of bank protection in the form of boulders and gabions. Flow is left to right.

Figure 3.7 Oblique view (June 1998) of the Chapel View field study reach showing the upstream most gravel bar form (PB1) and bank protection boulders (B). The flow is right to left.

Figure 3.8 Oblique view (June 1998) of a composite, erosive bank face in the Chapel View study reach. The bank face is approximately 1.5 m high.

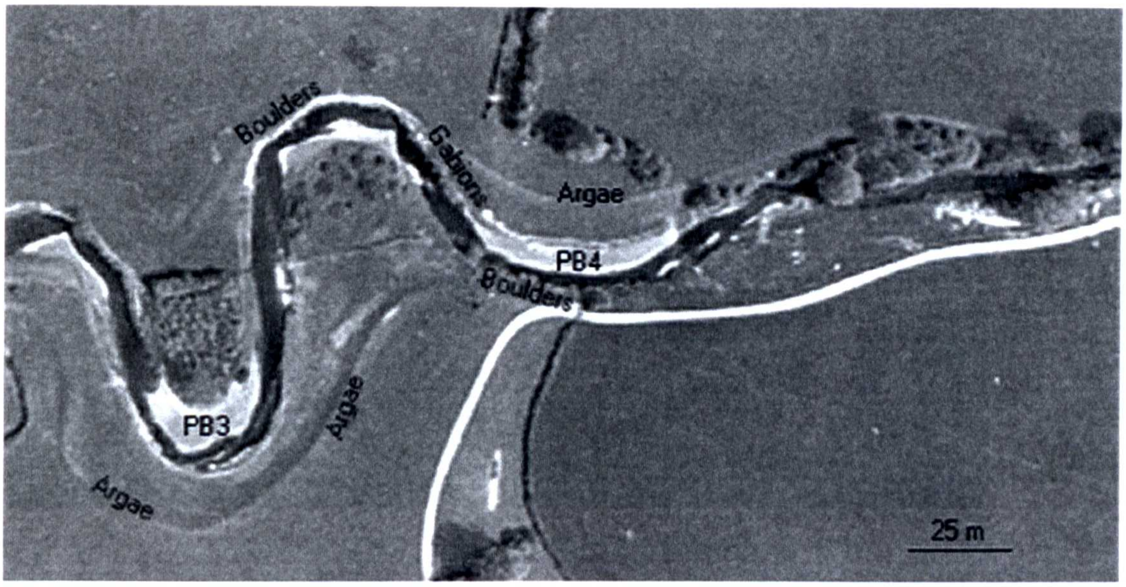


Figure 3.9 Aerial photograph (1995) of Bodiach Hall study reach showing positions of point bars (PB3 and PB4), argaes and remaining bank protection (boulders and gabions). Flow is left to right.

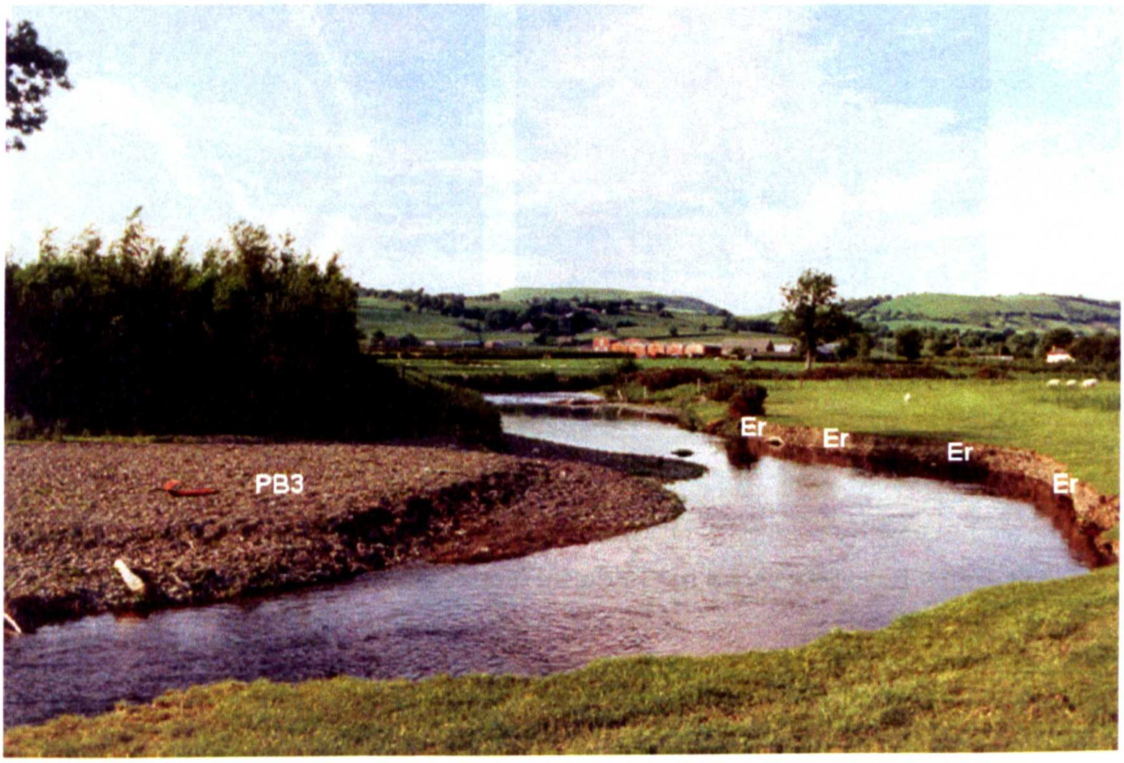


Figure 3.10 Oblique view of the upstream section of the Bodaich Hall reach (June 1998). The photograph was taken opposite the upstream most point bar (PB3) and shows the associated erosive face of the outer bank (Er). Flow is left to right.

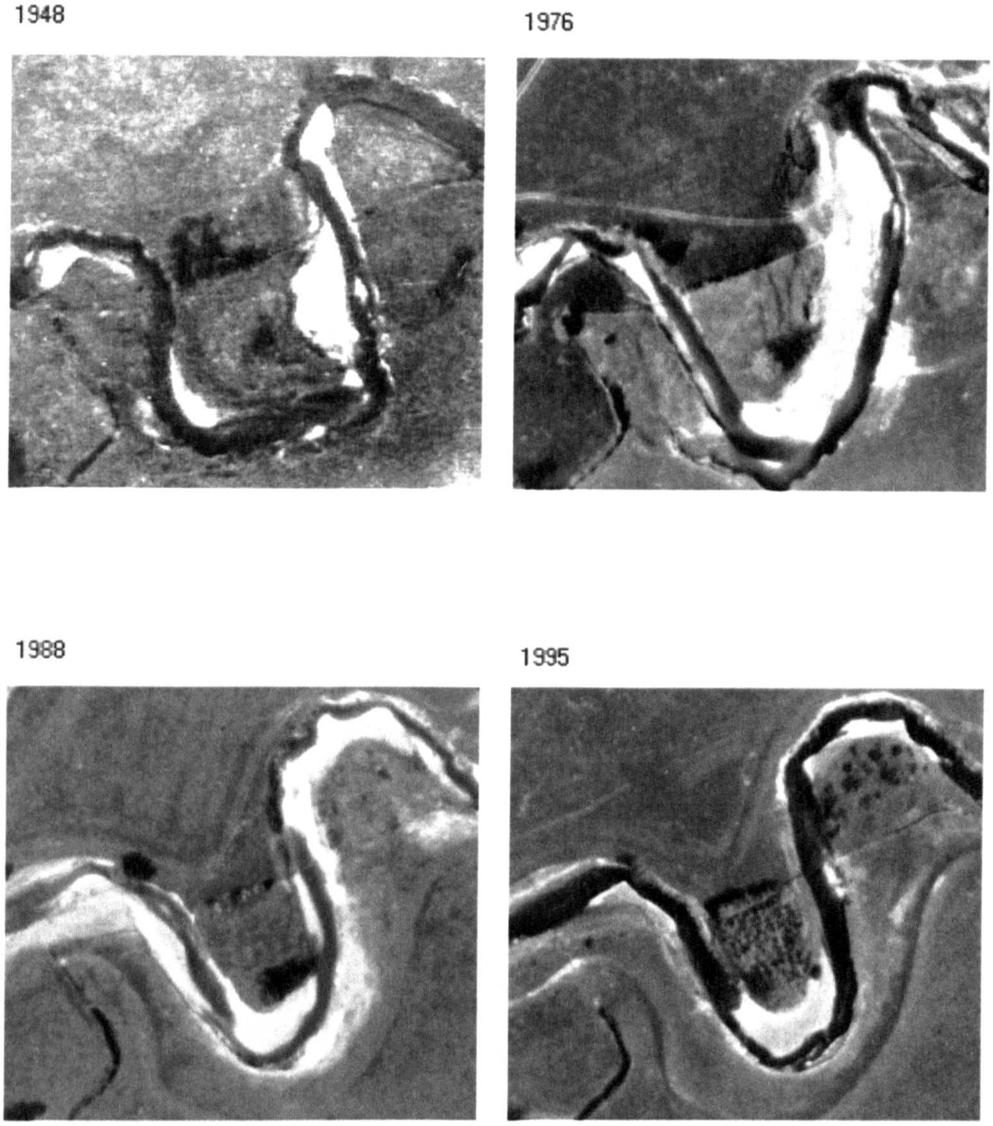


Figure 3.11 Example of medium-term lateral instability in a meander bend within the Bodiach Hall field study reach.

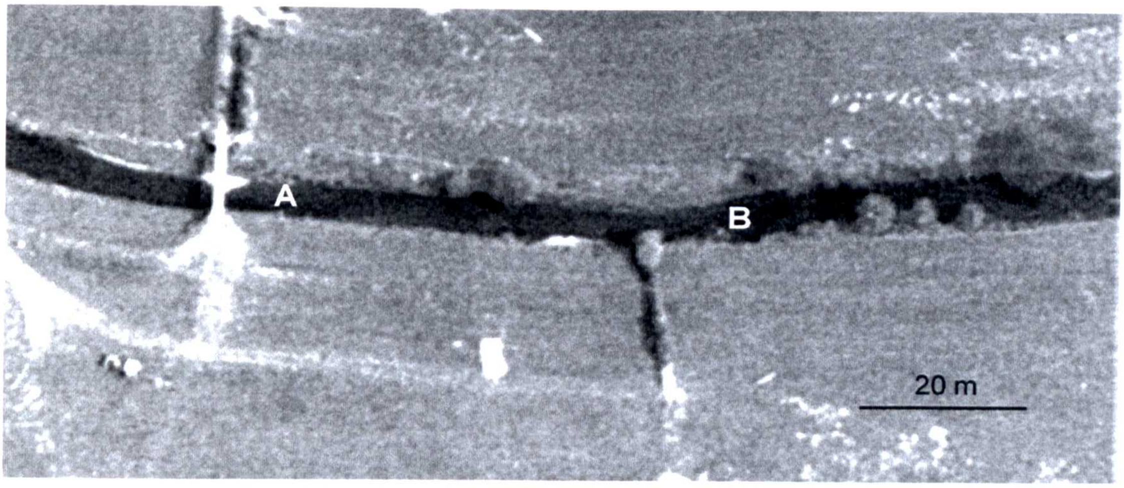


Figure 3.12 Aerial photograph of the Ddranen Ddu study reach (1995). The cross profiles are located between A and B. Flow is left to right.

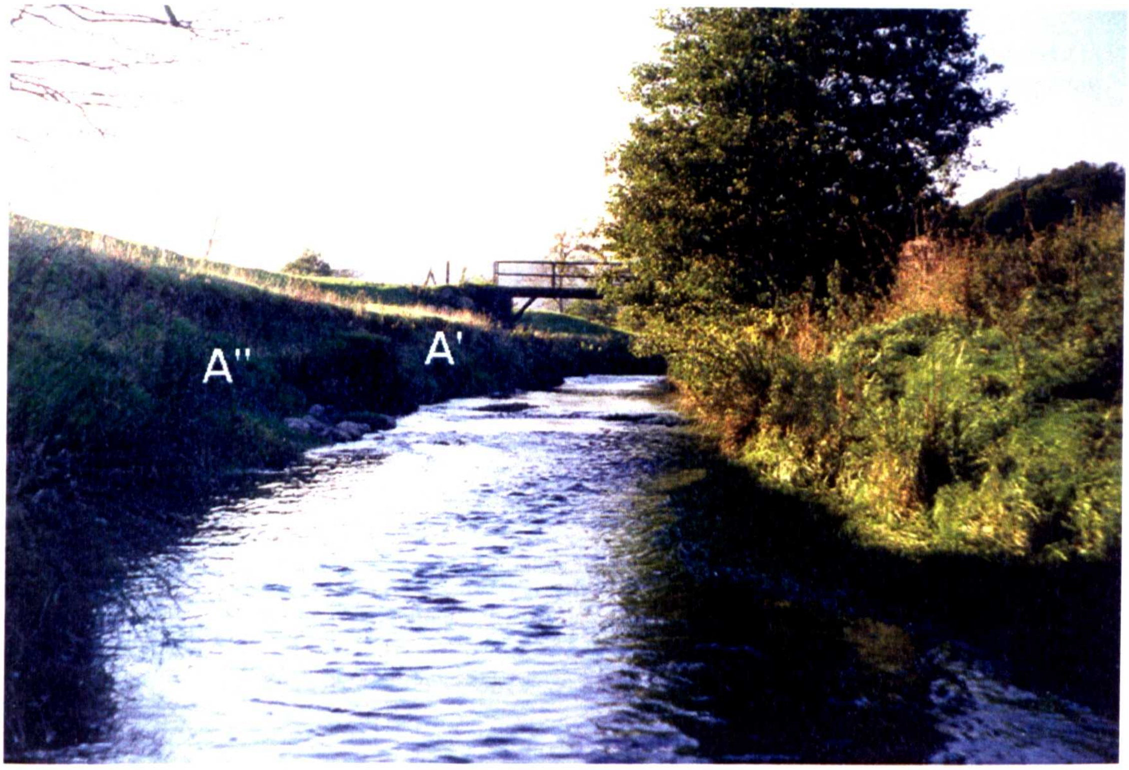


Figure 3.13 Oblique view (April 1998) of the Ddranen Ddu study reach looking upstream. The photograph was taken from the location of the downstream most cross profile. Evidence of rotational bank failure is visible at A' and A''.

3.6 Summary of techniques

The following tables (tables 3.2 – 3.5) summarise the major techniques used in the upper catchment and lowland study reaches and these are presented in the following four chapters, which address the project aims and objectives presented in section 3.1.

Table 3.2 Techniques presented in the following chapters used to satisfy the study objective 1.

Objective	Technique	Product
The generation of a medium-term upland catchment bed load yield for the Afon Trannon	• Digitisation of Tanllwyth and Trannon catchment parameters from O.S. 1:25000 scale maps	Formulae inputs for physiography / sediment delivery empirical formulae. Required to extrapolate Tanllwyth bed load yield to upper Trannon.
	• Digitisation of Dolgau forest plans	Quantification and analysis of forest cover. Required to provide % forest cover weighting to bed load yield extrapolation.
	• Mosaicing, rectification and digital terrain modelling of Tanllwyth and Upper Trannon catchments from air photos	Land surface slope analysis and comparison of the Tanllwyth and Upper Trannon catchments.

Table 3.3 Techniques presented in the following chapters used to satisfy the study objective 2.

Objective	Technique	Product
The generation of a medium-term hydrologic record for the Afon Trannon	• Data-based mechanistic rainfall – runoff modelling	A daily mean stage record for the Trannon at Caersws logger site covering 1969 to present.
	• Pressure transducer-type stage logging	A 15 minute and daily mean stage record for the Trannon at Trefeglwys logger site for 1999.
	• Bed load trapping	An estimate of the daily mean stage entrainment threshold of bed material in the Trannon at field study reach.
	• Magnetic bed load tracing	A further estimate of the daily mean stage entrainment threshold of bed material in the Trannon at field study reach.

Table 3.4 Techniques presented in the following chapters used to satisfy the study objective 3.

Objective	Technique	Product
The quantification of historical rates of channel change of the Trannon	• Raw aerial photo analysis of channel width and position in Paintshop Pro using photography from 1948, 1976, 1988 and 1995	A simple, quick and spatially extensive record of medium-term channel changes at the Trannon, however contains unquantified errors.
	• Geo correction and analysis of aerial photos for channel width and position using Ortho Max and ERDAS Imagine, using photography from 1948, 1963 and 1988	A record of channel changes at the Trannon . All errors are quantified and a medium quality output is produced.
	• Ortho rectification and analysis of aerial photography for channel width and position using Ortho Max and ERDAS Imagine, using photography from 1976 and 1995	A record of channel changes at the Trannon. A three dimensional output is possible with all error quantified.
	• Comparison of Paintshop Pro and ERDAS Imagine results	An assessment of the comparable accuracy of the photogrammetric techniques.
	• Digitisation of exposed gravel bars from rectified aerial photography	A record of the change in gravel bar size used tentatively to infer historic change in gravel storage.

Table 3.5 Techniques presented in the following chapters used to satisfy the study objective 4.

Objective	Technique	Product
A contemporary channel change record allowing examination of the processes of channel change	• The production of detailed, repeated, planform maps of field study reaches using total station surveys	A detailed record of short term changes in the planform of the lowland Trannon channel at selected study sites. The maps will allow the development of gravel bar forms and zones of active erosion to be monitored quantitatively and provide a data set which can be compared to channel changes identified from aerial photograph analysis.
	• Repeat cross profiling of benchmarked sections in the field study reaches	A fully quantified record of bed aggradation and degradation and bank erosion in the selected study reaches. Sections can be used to examine change in bed morphology at the meso-scale or greater, within channel sediment budgets and bed load transport rates (using inverse calculations).
	• Surface interpolation and digital terrain model construction of contemporary field survey data	Three dimensional record of channel morphology change in which changes in the volume of study reaches of selected regions of reaches can be visualised, quantified and examined. Essential for the easy identification of change in the submerged regions of the channel.
	• Detailed bank material mapping of Trannon from bed rock controlled reaches to below field study reaches	The identification of gravel and composite banks capable of yielding bed load to the channel as a result of bank erosion processes. When combined with historical bank erosion rates obtained from historical analysis will provide an estimate of bed material additions from main channel bank erosion. Can be used as a comparison to the estimated bed load yields from the upper catchment to infer relative importance of bed load sources.
	• Field identification of further sources of bed load within the catchment by tributary walking	The identification of any important bed load sources in the Trannon catchment overlooked by previous work or not accounted for by field and historical analysis.

4. UPPER CATCHMENT BED LOAD DELIVERY: MEDIUM-TERM BED LOAD YIELD ESTIMATE AND ANALYSIS

4.1 Introduction

The bed load yield of a catchment is defined as the total bed load outflow from a basin over a specified time period. It is normally highly variable between catchments and dependent upon many influencing factors. Traditionally, the problem of quantifying the sediment yield of a catchment has been approached in one of two ways:

1. Sediment routing models
2. Empirical formulae

Sediment routing models, whilst often extremely representative of the catchment, commonly have large numbers of input parameters. By mean of an example, a sediment routing model presented in Simons (1979) has 11 input parameters including physiographic, climatic and antecedent moisture parameters (see figure 4.1). The large number of parameters required in sediment routing models make them unfeasible for use in determining the bed load yield of the upper Trannon catchment, for which very little data exist.

Empirical formulae require little input information, and consequently provide the only feasible method for estimating sediment yields and delivery in the upper Trannon catchment. However, they have three main limitations:

1. Normally they allow an estimate of the sediment delivery rate (defined by Glymph (1954) as the ratio between the annual gross erosion in the catchment and the annual sediment yield) to be determined, rather than the sediment yield. Therefore, information about annual gross erosion in the catchment is required to generate a sediment yield from the sediment delivery rate.

Figure 4.1 Simons (1979) flow chart for a water and sediment routing model. Note the extensive input data required for parameters ranging from physiographic to climatic.

2. The empirical formulae are usually region specific. Consequently, if one wishes to apply empirical formulae to other catchments, one should select an empirical formulae derived for a region as physiographically and climatically similar as possible to the one of interest.
3. Empirical formulae are 'black box' in that they do not allow investigation into the processes of sediment delivery. Indeed, they assume that sediment delivery is a catchment property, whereas in truth it is also a channel property. This is especially true for bed load where bed load transport rates are highly dependent on discharge, channel parameters (i.e. depth and slope) and the size distribution of the bed material (Bagnold, 1980; Martin and Church, 2000).

There are a range of empirical sediment delivery formulae in the literature (Walling, 1993) which relate a number of input parameters to the sediment delivery rate. The input parameters can be grouped into two main categories:

- Catchment physiography parameters: Maner (1958), Roehl (1962), Williams and Berndt (1972), Williams (1977), Mutchler and Bowie (1975), Mou and Meng (1980)
- Catchment runoff: Mutchler and Bowie (1975)

It is clear from the above list that the relationship between catchment physiography and sediment yield has received the most attention in the literature. A range of physiographic parameters have been related to catchment sediment delivery rates. These are summarised in table 4.1.

Table 4.1 Examples of relationships between sediment delivery rates and catchment physiographic characteristics

Author / date	Study location	Equation
Maner (1958)	Kansas, USA	$\log DR_e$=2.94259-0.82363 colog R/L
Roehl (1962)	Southeastern USA	$\log DR_e$=2.88753-83291 colog R/L
Williams and Berndt (1972)	Bushy Creek, Texas, USA	$DR_e=0.627\ \mathrm{SLP}^{0.403}$
Williams (1977)	Texas, USA	$DR_e=1.366\times10^{-11}A^{-0.100}R/L^{0.363}CN^{5.444}$
Mou and Meng (1980)	Dali River, China	$DR_e=1.29+1.37\ln Rc-0.025\ln A$

DR_e = Estimated sediment delivery rate

R/L = catchment relief to length ratio. Relief is defined as the difference between the mean elevation of the watershed divide at the headwaters of the main stem drainage, and the elevation of the stream bed at the point of sediment yield measurement. Length is defined as the distance, measured essentially parallel to the main stream channel, from the point of sediment yield measurement to the watershed divide.

SLP = main channel slope

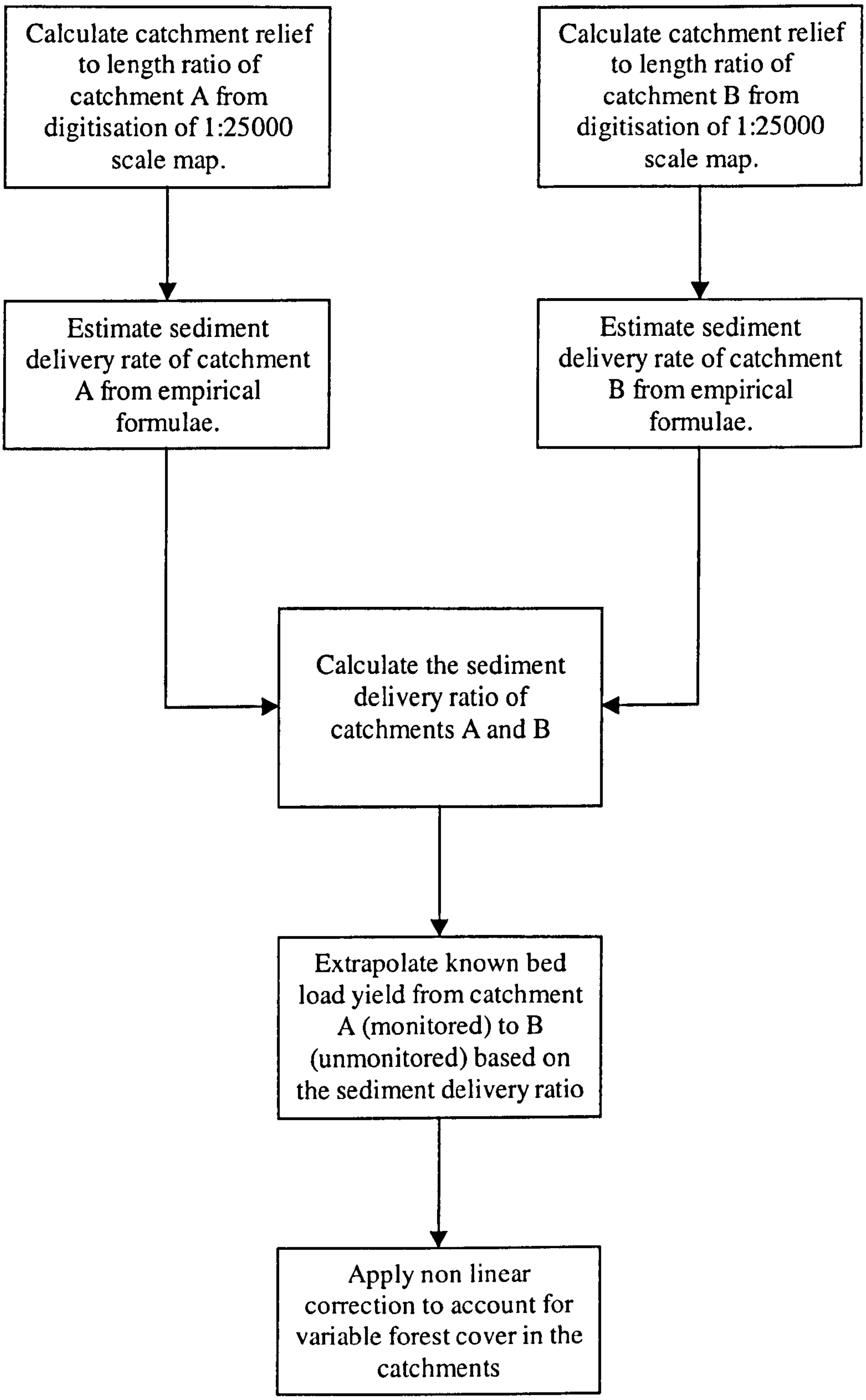
A = catchment area

CN = US Dept. of Agriculture Soil Conservation Service curve number

Rc = gully density

Of the above examples, Mou and Meng (1980) is primarily concerned with sediment delivery from gully erosion. As such it is not representative of processes in upland UK forested catchments where gullies are seldom cited as major sediment producers. Williams (1977) requires soil information which is not available in the upper Trannon. Williams and Berndt (1972) provides a river specific relationship for the arid state of Texas. Therefore its ability to represent sediment delivery from forested

Figure 4.2 Flow diagram showing the procedures used to generate medium-term upper catchment bed load yields for the Afon Trannon.



upland UK catchments is extremely suspect. Maner (1958) and Roehl (1962) however, successfully provide relationships which have been tested on:

- a large number of catchments: 25 and 38 respectively
- catchments of varying size: minimum 0.036 - maximum 322 sq. mi.
- catchments in various geographical locations: ranging more than 1000 miles from the central southern state of Texas to the eastern state of North Carolina
- catchments with varying physiography: from the Blue Ridge Mountains to the Southern Coastal Plains

Consequently, of all the available relationships the Maner and Roehl formulae are selected as being most likely to provide a realistic sediment delivery rate for the upper Trannon catchment due to their versatility. Maner (1958) calculates that the R/L ratio accounts for 97 % of the variation in sediment delivery in the 25 catchments whilst Roehl (1962) demonstrates that the R/L ratio accounts for 88 % in the slightly more variable 38 catchments studied.

In order to determine the sediment yield of the upper Trannon catchment, it is necessary to have information pertaining to gross erosion of bed load sized material in the upper catchment. Unfortunately this information is unavailable. However, if the sediment delivery rate of the Trannon catchment and a further catchment, for which the sediment yield and sediment delivery rate is known, are compared, it is possible to extrapolate the sediment yield data from the known catchment to the unknown based on the difference between their respective sediment delivery rates (here termed the sediment delivery ratio). The procedure is summarised in figure 4.2.

Newson (1980) used such a catchment comparison for the Wye and Tanllwyth catchments, mid-Wales, to examine whether the elevated bed load yields in the forested Tanllwyth catchment were a result of differences in the catchment physiography of the two catchments rather than a result of forestry in the Tanllwyth.

He used the formulae of both Maner (1958) and Roehl (1962), and assumed that the gross erosion of source areas was the same in both catchments. This assumption, however, remained untested. Physiographic parameters for the two catchments are presented in table 4.2.

Table 4.2 Physiographic characteristics of the Cyff and Tanllwyth catchments

		Cyff	Tanllwyth
Area		3.13 km^2	0.89 km^2
Channel gradient		58.2 m km^{-1}	79.4 m km^{-1}
Overland slopes 0-9°		47.5 %	69.4 %
(% area)	10-19°	52.5 %	30.6 %
Drainage density		1.90 km km^{-2}	3.30 km km^{-2} (including drains)
Mainstream length		2.90 km	1.50 km

From the above table it would be expected that the smaller catchment area, steeper catchment slopes, shorter channel length and higher drainage density in the Tanllwyth would translate to a higher bed load delivery rate for the Tanllwyth compared to the Cyff. Indeed, the Maner and Roehl sediment delivery rate estimates confirmed this with the sediment delivery ratio between the two catchments estimated at between 1.2:1 and 1.4:1. The measured bed load delivery ratio for the two catchments was 3.7:1 (Newson 1980) strongly suggesting that the forestry in the Tanllwyth, rather than just the difference in physiography, was responsible for elevated bed load yields. However, Newson (1980) notes that the Maner and Roehl formulae were generated using sediment yield data collected from sediment deposition in lakes and reservoirs. As such, the formulae relate to the delivery of both suspended and bed load combined rather than solely bed load. Because suspended sediment is transported through catchments considerably faster than bed load, one might reasonably expect the Maner and Roehl formulae to overestimate the bed load delivery rate for a catchment.

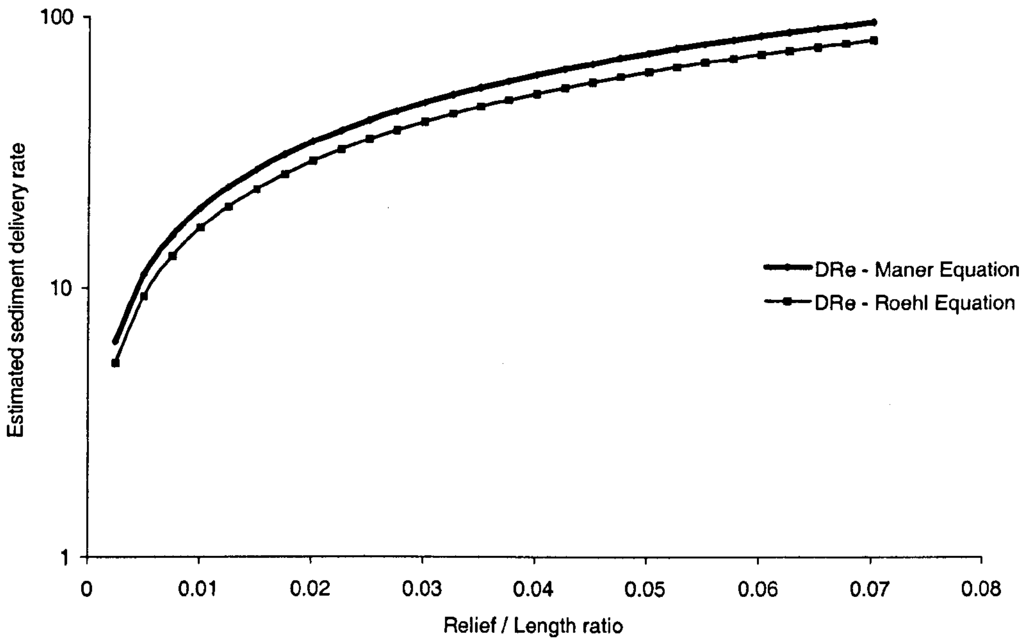


Figure 4.3 Comparison of estimated sediment delivery rates for a range of catchment relief to length ratios, computed using the formulae of Maner (1958) and Roehl (1962).

In this chapter the procedures and results used in the estimation of a bed load yield for the upper Trannon catchment are presented and discussed. The bed load delivery rate is estimated for the Tanllwyth catchment (which lies only 5 km to the southwest of the upper Trannon) and the upper Trannon, using the Maner and Roehl formulae. Subsequently, the bed load delivery ratio for the two catchments in estimated and the measured bed load yield for the Tanllwyth catchment is extrapolated to the upper Trannon based on this ratio. Finally, a correction for the lower % forest cover in the upper Trannon is made.

4.2 Estimating medium-term sediment yields

4.2.1 Determination of bed load yield from estimated sediment delivery rates

The Maner (1958) and Roehl (1962) formulae are presented below. Both formulae (eqn. 4.1 and 4.2) are extremely similar (also see figure 4.3) with only small differences in the coefficients:

Maner (1958): $\log DR_e = 2.94259 - 0.82363 \operatorname{colog} R/L$ (equation 4.1)

Roehl (1962): $\log DR_e = 2.88753 - 0.83291 \operatorname{colog} R/L$ (equation 4.2)

Where: DR_e is the estimated sediment delivery rate

R/L is the catchment relief to length ratio

Sediment delivery rates were calculated for the Tanllwyth and Trannon catchments using both the Maner and Roehl formulae. R/L ratios were determined for the Tanllwyth and upper Trannon catchments from the 1:25000 scale O.S. map. The ratios and subsequent calculated sediment delivery rates are presented in table 4.3. The estimated delivery rates vary in magnitude by up to 14% depending on the equation used in the calculation, however the ratio of DR_e between the Tanllwyth and upper Trannon catchments is almost identical for both equations; unsurprising given the similarity of equations 4.1 and 4.2. It is estimated that, in conditions of identical sediment supply per unit area, the Tanllwyth delivers between 1.62 and 1.64 times the sediment delivered by the Trannon to the catchment outflow.

Assuming equal gross erosion in both catchments, and using a DR_e ratio of 1.63, the upper Trannon catchment mean annual bed load yield is estimated as 25.33 t km^{-2} yr^{-1}. Although difficult to fully justify without field data, the assumption of equal gross erosion is considered valid based on the similarity in forest practices within the two catchments and their close geographical proximity. Both the Tanllwyth and upper Trannon catchments were forested in the late 1940's, and, in the main part, by the same organisation (The Forestry Commission). As a result, similar preparation and ditching practices were employed in both catchments, with a herring bone ditch pattern being established at an approximate spacing of 20 m apart. Because it is the erosion of these ditches which provides the majority of the bed load yield to UK forested streams, the similar ditch configuration in the two catchments would imply a very similar erosion source area. Ditch erosion is strongly related to rainfall patterns (see section 2.1.2.1), therefore, it would seem feasible that as long as rainfall within the two catchments was similar, gross erosion of the ditches would also be similar. This is likely given the close proximity of the two catchments.

Table 4.3 *R/L* ratios for Tanllwyth and upper Trannon catchments and estimated sediment delivery rates.

Catchment	***R/L* ratio**	**DR_e (Maner eqn.)**	**DR_e (Roehl eqn.)**
Upper Trannon	0.0375	59%	51%
Tanllwyth	0.07	97%	83%
DR_e ratio		1.64	1.62

Table 4.4 Bed load yields from the 0.9 km^2 Tanllwyth catchment (1973 – 1980). Source: Sawyer, A. *pers. comm.* Note the annual variability in the yield which is most likely due to stream flow conditions, particularly the influence of storms.

Year	Bed load yield (tonnes km^{-2})
1973	77.53
1974	25.17
1975	22.36
1976	11.69
1977	84.94
1978	21.12
1979	70.67
1980	16.48
MEAN	**41.29**

4.2.2 Accounting for the impact of variable percentage forest cover

The Tanllwyth and upper Trannon catchments have differing percentage forest cover. The Tanllwyth is entirely forested, whereas the upper Trannon currently has 56 percent forest cover. This difference has important implications for bed load yields within the catchments because forest drainage ditch erosion forms the major bed load source. A lower percentage forest cover would infer fewer ditches, less gross ditch erosion per unit area and, hence, a lower bed load yield. Stott (1997a) investigated the relationship between percentage forest cover and bed load yields. He was able to demonstrate that the percentage of the catchment forested is indeed an important parameter in the volumes of bed load yielded from an upland afforested catchment. In his study he compares data from eleven forested catchments in England, Wales and Scotland, in which bed load yields have been monitored and forest cover is variable. The relationship he identifies is shown in figure 4.4. The

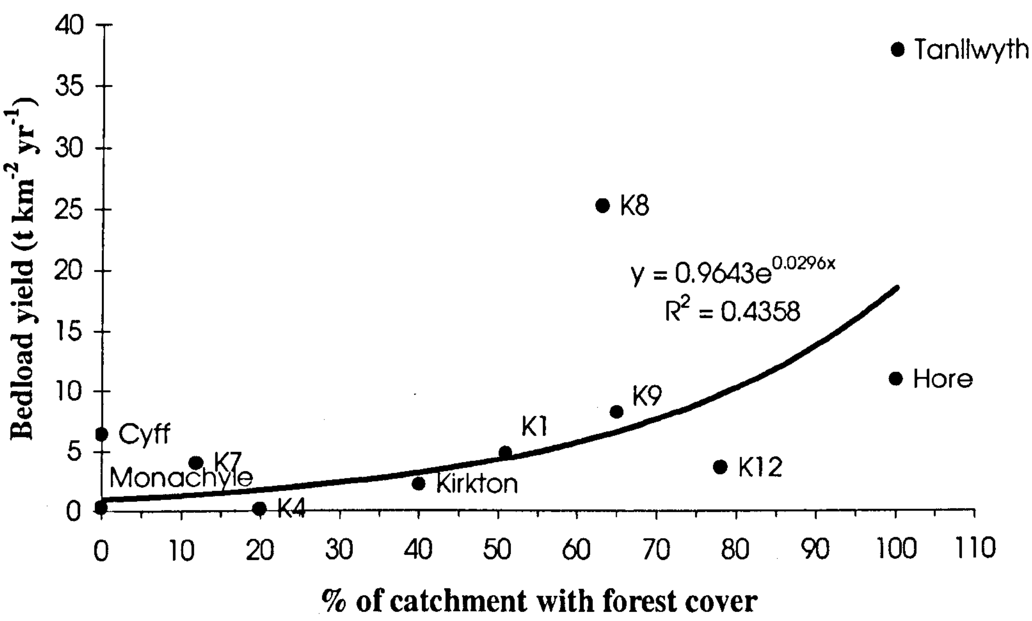


Figure 4.4 Scattergraph of subcatchment bed load yield versus percentage catchment forest cover. Graph shows data from Kirkton glen, Balquhidder, Scotland with the Kirkton (K1-K12) and Monachyle yields (Stott, 1987; Johnson, 1993) and the Hore, Wales (Leeks, 1992). The Cyff and Tanllwyth catchments at Plynlimon, mid-Wales (Moore and Newson, 1986) are also shown. From Stott (1997a).

scatter plot of the data reveals an exponential increase in bed load yield (y) with increasing forest area (x):

$$y = 0.9643e^{0.0296x} \qquad \text{(equation 4.3)}$$

However, the strength of the relationship is statistically weak, particularly for catchments where the percentage forest cover is high. The line of best fit has an R^2 of 0.4358 with residual values becoming greater with increasing forest cover. Both the residuals for the Tanllwyth and Kirkton 8 bed load yields are considerably higher than the rest of the data set. The relationship also fails to take account of the varying physiographic, climatic and geologic settings of the catchments which are undoubtedly important parameters which influence bed load yields.

When applied to the Tanllwyth catchment, with 100 % forest cover, the relationship predicts a bed load yield of 19.29 t km^{-2} yr^{-1}. This is a 53 % underestimation of the true Tanllwyth mean annual bed load yield. Applied to the upper Trannon catchment, in its current state of 56 % forest cover, the relationship predicts a bed load yield of 6.35 t km^{-2} yr^{-1}. The implication, therefore, is that the previously estimated bed load yield, achieved by direct catchment extrapolation, and taking no account of forest cover, may be a minimum of 3.7 times too great.

Figure 4.5 presents the medium-term percentage catchment forest cover of the upper Trannon. It can be seen that forest cover has increased progressively since 1949, when forest cover was 12 %, to 1978 when forest cover stabilised at the current level of 56 %. This progressive increase means that simply dividing the annual bed load yield by 3.7 to account for forest cover is invalid in the period prior to 1978, because percentage forest cover in the upper Trannon was significantly lower than current levels. Therefore extrapolated bed load yields have been weighted year by year to account for the impact of variable forest cover. The non-linear weighting used is a direct application of Stott's relationship. The final extrapolated bed load yields, accounting for variation in forest cover, are presented cumulatively in figure 4.6. It

FIGURE 4.5 AND 4.6

Figure 4.5 Percentage of upper Trannon catchment forested between 1949 and 1999. It should be noted that although no further forest area increase occurred after 1978, felling and second rotation forestry was extensive.

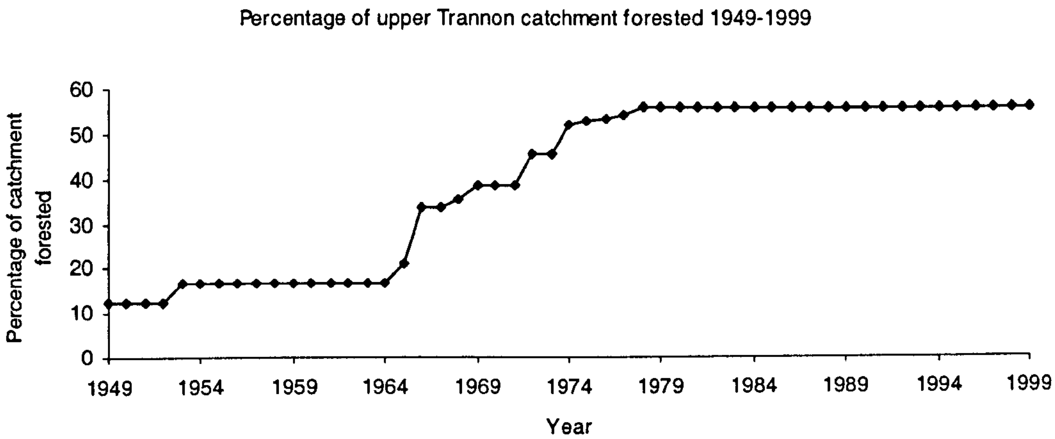
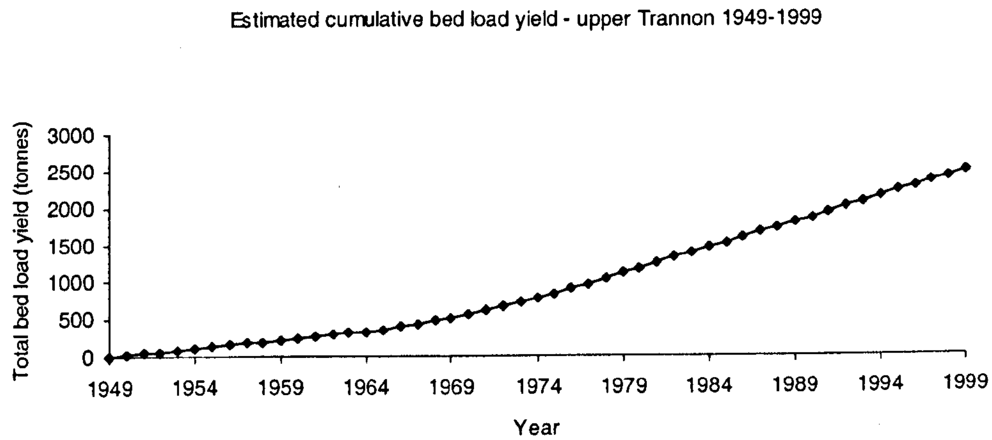


Figure 4.6 Estimated cumulative bed load yield for the upper Trannon 1949-1999.



is estimated that 2490 tonnes of bed load have been delivered to the Trannon from the upper catchment since 1949. In the period prior to 1974 it can be seen that cumulative estimated bed load yields are very low. They total only 30 % of the 1949-1999 total yield in 50 % of the study period. This demonstrates the important influence of the low percentage forest cover in the catchment between 1949 and 1965.

4.2.2.1 Bed load mass to volume conversion

Throughout this research bed load bulk density is assumed to equal 2650 kg m^{-3} and bed porosity is assumed to equal 30 % (Ferguson, *Pers. Comm.*). Therefore, accounting for porosity, the density of the bed is:

$$\text{Bed density} = 2650 \text{ kg m}^{-3} \times 0.7 = 1855 \text{ kg m}^{-3} \qquad \text{(equation 4.4)}$$

Using the above conversion, the medium-term bed load yield of the upland Trannon catchment can be converted from 2490 t to 1342.3 m^3.

4.2.3 The Tanllwyth and upper Trannon catchments: further comparison

The formulae of both Maner and Roehl provide estimates of catchment bed load yields with knowledge of the catchment relief to length ratio only. Despite the reportedly good results obtained by using solely this parameter, it seems unlikely that other factors are not also influential in UK catchments. Both climate and catchment geology are intuitively important. Geology is commonly an overriding control of bed load availability and catchment climate, particularly rainfall pattern and magnitude, determine bed load transport within the catchment. The upper Trannon and Tanllwyth catchments have an extremely similar geologic setting, being a mixture of friable Palaeozoic grits, mudstones and shales, or drifts composed of them (Rudeforth, 1967). The thickness of the till drift is spatially variable and overlain by peat and stagnopodzols in both catchments (Soil Survey of England and

Wales, sheet 2, 1:250,000). It is the drift which forms the major bed load source in both catchments. Similarly, the close proximity of the two catchments means that rainfall patterns and magnitudes are similar. Consequently, variance in the catchment geology and climate between the Tanllwyth and Trannon is considered insignificant.

Catchment physiography is not fully described by the relief to length ratio of Schumm (1954). Therefore, in order to provide a more comprehensive catchment comparison, eight physiographic parameters, identified by Newson (1980) as affecting gross erosion rates and sediment yields, have been measured via digitization of the O.S. 1:25000 map and are presented in table 4.5. Slope angles presented within the table have been calculated using the DTM construction procedures described in section 4.2.4.1.

Table 4.5 Major physiographic characteristics of the Tanllwyth and upper Trannon catchments influencing gross erosion rates and sediment yields.

	Tanllwyth	**Upper Trannon**
Catchment area	0.9 km^2	10.8 km^2
% of catchment forested	100%	56%
R/L ratio	0.07	0.0375
Within-forest channel gradient	79.4 m km^{-1}	49.9 m km^{-1}
% forest area with slope 0-9°	41%	62%
% forest area with slope > 10°	59%	38%
Forest drainage density	2.5 km km^{-2}	Data not available
Within-forest main channel length	1.5 km	3.4 km

Percentage forest cover and relief to length ratio have already been accounted for in the previous paragraphs. The higher relief to length ratio of the Tanllwyth is reflected in its higher catchment slope angles and within forest channel gradient. Therefore, one can argue that use of the relief to length ratio accounts for much of the variability in these parameters. What it does not account for, however, is the large variation in size between the Tanllwyth and upper Trannon catchments, a factor that is also reflected in the longer within-forest channel length of the upper Trannon.

This increased catchment size has strong implications for within-forest bed load storage and hence the lag time existing between material being eroded from the drainage ditches and exiting the upper catchment. The larger size of the upper Trannon would infer greater lag times, in particular opportunities for bed load storage in the main channel could enhance the lag time significantly. The lag time has not been accommodated in the procedures presented in the previous paragraphs and it is therefore extremely likely that the upper Trannon estimated cumulative bed load yield for the period 1949-1999 is high.

4.2.4 Investigation of forest slope angles

Newson (1980) demonstrates the overriding importance of forest drainage ditch erosion as the source of enhanced bed load yields in afforested Welsh catchments. Where slopes are steepest the potential for ditch erosion is enhanced increasing gross unit erosion. Examination of the 1:25000 O.S. map suggests considerably higher slope values in the Tanllwyth catchment. As a result within-forest slope angles for the Tanllwyth and upper Trannon catchments have also been examined in detail.

4.2.4.1 Quantifying catchment slope angles from digital terrain models: Procedure

1995 stereo pair photography at approximately 1:20000 scale was available for use in this study; covering 100% of Dolgau Forest (or 56 % of the upper Trannon catchment) and 60% of the Tanllwyth catchment. This allowed digital terrain models (DTMs) to be constructed for both catchments within a GIS, and topographic parameters to subsequently be quantified.

Photographs were placed on a UMAX 1220p scanner (capable of 1200 × 600 dots per inch (dpi) optical resolution, and aligned so that the fiducial marks were within the boundaries of the scanning bed. Images were scanned at 600 × 600 dpi as black and white photographs so that each pixel represented a square on the image 42 μm × 42 μm. This allowed a ground resolution of 0.82 m to be obtained from the 1:20000 scale photography; in other words, each pixel to which a grey scale was assigned covered a ground area of 0.82 m^2.

Images were mosaiced and colour adjusted and DTMs extracted in OrthoBASE. In excess of 20 ground control points and a subsequent 30 tie points were used to form the Trannon upper catchment model and 12 ground control points and a further 18 tie points were used for the smaller Tanllwyth model. OrthoBASE extracts DTMs using pattern recognition algorithms (in the form of collinearity equations) to match a stereo-pair set of images and calculate surface elevation. It was decided that a DTM with a ground resolution of 5 m × 5 m was acceptable for extracting catchment slope information. A higher resolution DTM may be error prone due to the fact that the surface being modelled was in fact the irregular surface of the forest canopy and not the true ground surface. As a final ground resolution of only 5 m × 5 m required, ground control provided by the O.S. 1:25000 scale map was considered able to provide acceptable accuracy. Root mean square error of the ground control was <5 m suggesting that the ground elevation and position extracted from any point on the model was absolutely accurate to a sub 5 m accuracy although relative accuracy across the DTM was < 1m. DTMs were cropped to the catchment watershed prior to slope analysis.

Slope values were extracted from the DTM in ERDAS Imagine using the 'surface slope' function. The function yields slope angle data direct from the interpolated surface grid. The slope angle information is stored as a data array which can be exported. The output from ERDAS Imagine was imported into Excel allowing summary statistics to be produced.

4.2.4.2 Quantifying catchment slope angles from digital terrain models: results and canopy margin corrections

DTM models of densely forested areas represent the surface created by the canopy and not the ground surface. In coniferous plantations, where tree age is the same throughout large forestry parcels the canopy surface adequately represents the ground surface. However, at the margins of a forest parcel a height discrepancy of several meters exists. The surface interpolation algorithms cause a ridge to form in the DTM along these margins which are identifiable due to their extremely high slope angles. Figures 4.7 and 4.8 show slope angles within Dolgau forest and the Tanllwyth catchment as determined from the DTMs. Lines of pixels with slope values >30 degrees (coloured red) are clearly visible along the margins of forest parcels where trees are juxtaposed against forest tracks. As a result canopy height discrepancy exists between the trees and the road, and the discrepancy becomes more pronounced the greater the age of the forest parcel. Ground truthing of both catchments together with examination of the 1:25000 O.S. maps of the areas suggested that slope angles were seldom greater than 40 degrees, so to remove the canopy margin error slope values of > 40 degrees were ignored in the slope analysis.

Summary statistics from the catchment DTMs are presented in figure 4.9. The data trend is similar for the Tanllwyth catchment and Dolgau forest, although on the whole the Tanllwyth is the steeper of the two catchments. This would infer that the Tanllwyth has a greater potential to deliver bed load to the main channel than the Trannon, particularly when the impact of the smaller catchment size previously discussed is considered. It should, however, be remembered that the data for the Tanllwyth represents only the upper 60 % of the catchment where slope angles are steepest. It is possible that analysis including the less steep slopes of the lower 40% of the catchment would show only minor differences in the slope statistics for the two catchments.

Figure 4.7 Stratified digital terrain model of Dolgau forest in the upper Trannon catchment (scale 1:25000). The numbers on the key relate to slope angle in degrees.

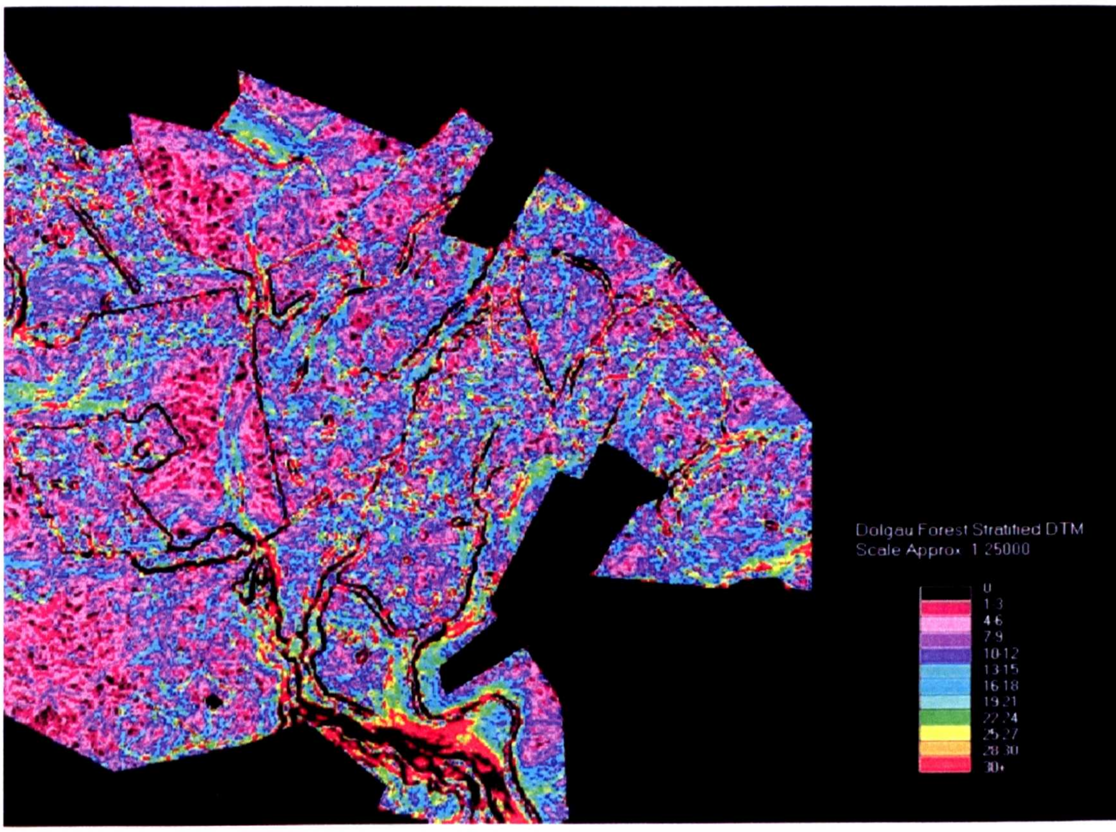


Figure 4.8 Stratified digital terrain model (scale 1:3750) of the Tanllwyth catchment.

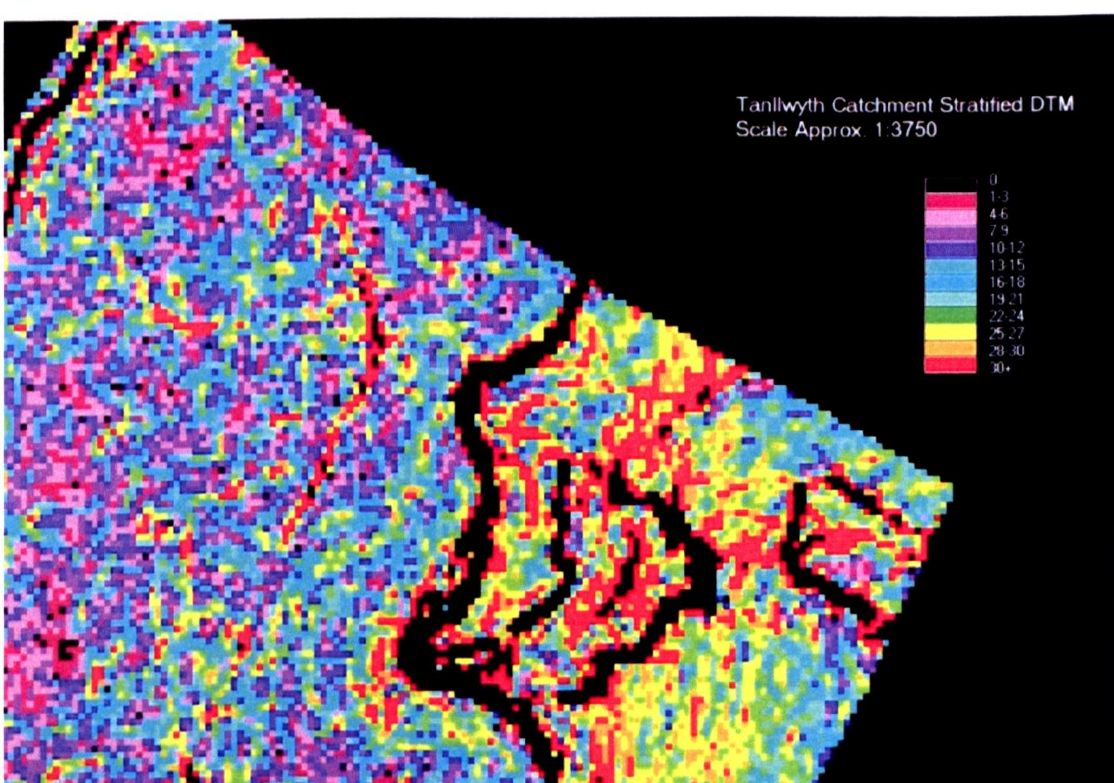
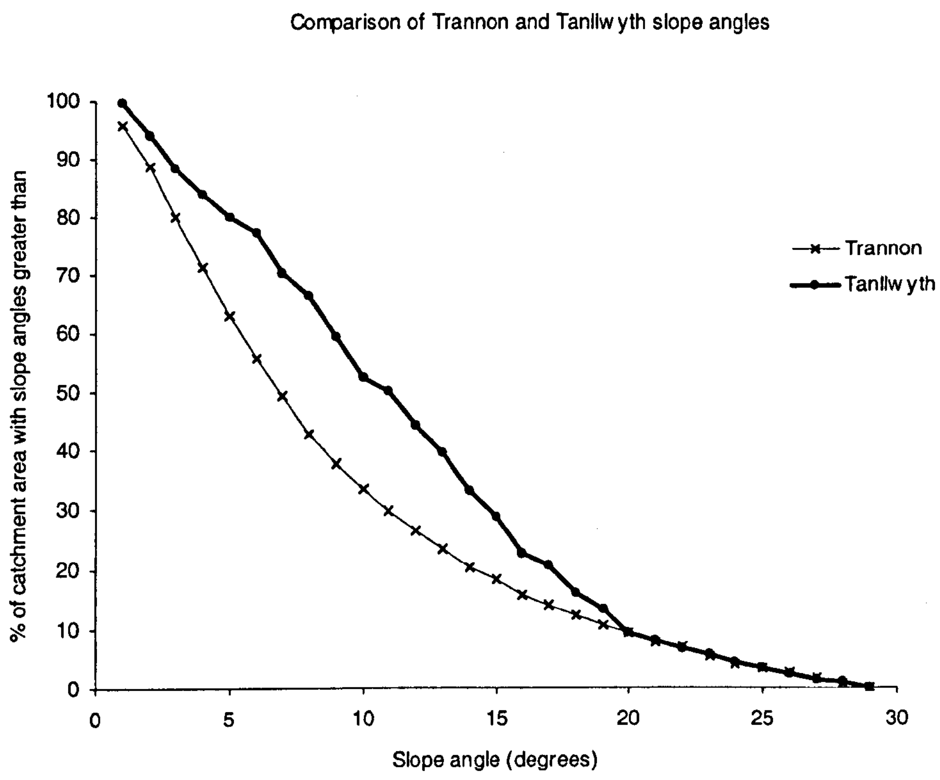


Figure 4.9 Inverse cumulative comparison of Dolgau forest (upper Trannon) and Tanllwyth catchment slope angles extracted from catchment DTMs.



4.3 Comparison with other UK upland catchments

With the current catchment setting of 56 % forest cover, a bed load yield of 6.35 t $km^{-2}\ yr^{-1}$ is estimated for the upper Trannon catchment, equating to an annual total catchment bed load yield of 68.58 t yr^{-1}. This figure incorporates the effects of both catchment *R/L* ratio and % forest cover. However, lower forest cover prior to 1978 is reflected in far lower estimated bed load yields. In total it is estimated that 1342.3 m^3 of bed load have been yielded from the upper Trannon catchment in the period 1949-1999. Investigation of forest slope angles implies a higher availability of sediment from the Tanllwyth and the smaller catchment size would imply shorter lag times between sediment erosion and discharge from the catchment. No field evidence is available to directly support these views, but they do provide limited circumstantial evidence that the estimated bed load yields for the upper Trannon catchment, although probably accurate to within an order of magnitude, may be rather high.

As stated earlier, the methodology employed here to generate a bed load yield estimate for the upper Trannon assumes that bed load yield is a catchment physiography property as is the case for suspended sediment for which the sediment delivery formulae were originally derived. However, unlike suspended sediment bed load yield is normally transport limited in UK catchments, therefore it is really controlled by a combination of catchment physiography properties and channel variables. The channel control becomes particularly dominant if one is interested in the investigation of bed load yield at high temporal resolutions (i.e. the single event scale).

In order to investigate the extent to which a relationship exists between catchment relief to length ratio and medium-term bed load yield, data from a range of monitored UK catchments have been compared (table 4.6). A scatter plot of catchment relief to length ratio versus bed load yield is presented in figure 4.10. The

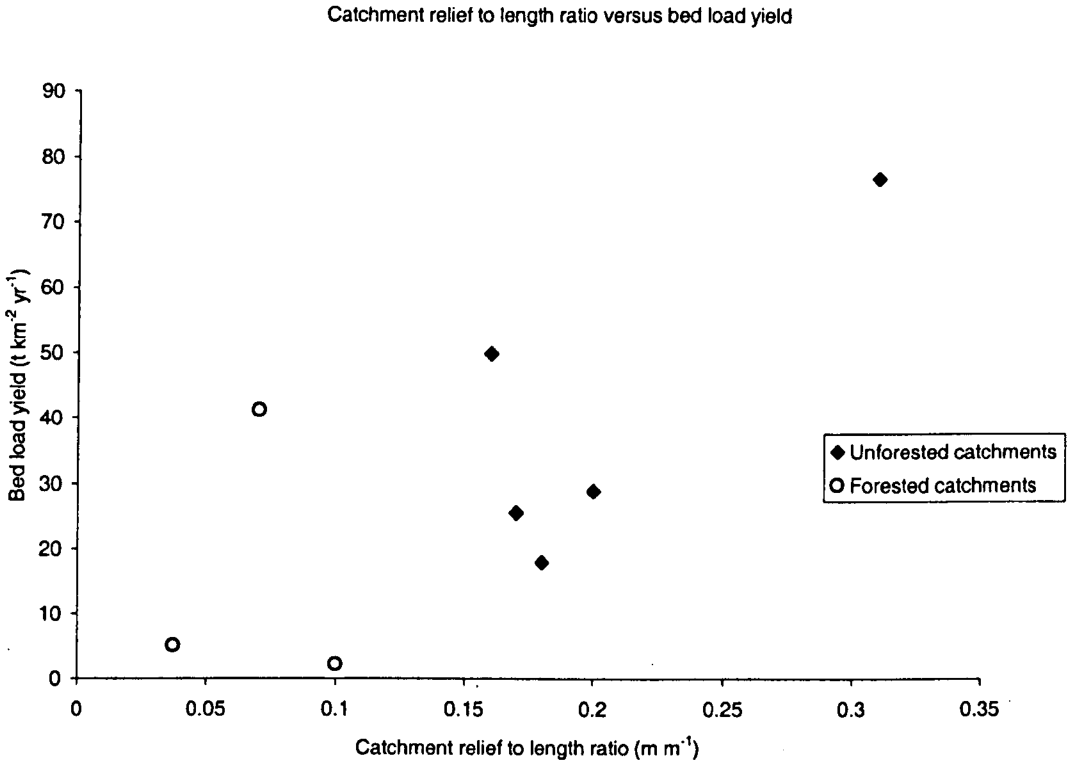


Figure 4.10 Mean annual bed load yield and catchment relief to length ratio for eight upland catchments in the UK: Trannon, mid-Wales (this study), Tanllwyth, mid-Wales (Newson, 1980), Hore, mid-Wales (Leeks, 1992), Kirkton Glen, Scotland (Johnson, 1993), Allt Daim, Scotland (Richards and McCaig, 1985), Beckthorn, England (Newson and Leeks, 1985), Coledale, England (Newson and Leeks, 1985) and Lanthwaite, England (Newson and Leeks, 1985).

data show a positive relationship between catchment relief and bed load yield with steeper catchments having a higher bed load yield and suggests that bed load yield and catchment physiography are indeed linked. However, the residual scatter is also large and this is very probably due to the influence of channel controls.

Table 4.6 Comparison of upland catchment bed load yields in the UK.

Study	**Catchment area (km^2)**	**Relief: length ($m\ m^{-1}$)**	**Forest cover?**	**Bed load yield ($t\ km^2\ yr^{-1}$)**
Afon Trannon, mid-Wales	10.8	0.0375	Yes (56 %)	6.35
Tanllwyth, mid-Wales (Newson, 1980)	0.9	0.07	Yes (100 %)	41.29
Hore, mid-Wales (Leeks, 1992)	3.08	NA	Yes (100 %)	11.8
Kirkton, Scotland (Johnson, 1993)	6.85	approx. 0.1	Yes (40 %)	2.2
Allt Daim, Scotland (Richards and McCaig, 1985)	5.82	0.18	No	18
Allt a'Mhuillin, Scotland (Richards and McCaig, 1985)	6.19	0.17	No	25.7
Beckthorn, England (Newson and Leeks (1985)	0.5	0.31	No	77
Coledale, England (Newson and Leeks, 1985)	6.0	0.16	No	50
Lanthwaite, England (Newson and Leeks, 1985)	4.0	0.2	No	29

5. THE TRANNON HYDROLOGIC SETTING: MEDIUM – TERM RECORD GENERATION AND ANALYSIS

5.1 Rainfall – runoff modelling

5.1.1 Introduction

Medium-term investigations into the causes of lowland river channel destabilisation necessarily require knowledge of both the sediment and hydrologic dynamics (in particular the frequency with which flow capable of bed load transport occurs) of the river throughout the study period. In the case of the Afon Trannon hydrologic data, consisting of a 15 minute stage record, exist for the lowland channel for the period April 1994 to February 2000. Prior to 1994 the river was not gauged. Therefore there is a need to generate a historical hydrologic record for the lowland channel. A historical (1969 – 2000) daily total rainfall record exists for a monitoring site at Dollyd in the Severn catchment, some 5 km west of the upper Trannon catchment. Consequently, the potential of using the Dollyd rainfall record and the limited lowland channel stage record as inputs in a rainfall-runoff model were examined.

Rainfall is converted to runoff via an array of physical catchment processes which interact to form a complex system. The internal structure of interactions in the system filters the rainfall input into the flow (or in this case stage) output; therefore the aim of rainfall – runoff models is to accurately represent the filter:

$$\text{Input} \longrightarrow \boxed{\text{FILTER}} \longrightarrow \text{Output}$$

Rainfall – runoff models attempt to model the filter in one of three main ways:

1. Deterministically – (e.g. Stanford Watershed Model) where the internal mechanism of the filter is determined and the influential parameters are

identified. Accordant parameter data sets can then be collected and converted to an output via algorithms designed to be representative of the internal mechanism.

2. Data-based – where the mathematical relationship between the input and output data sets is identified and applied to the input, thereby generating an output.
3. Data-based mechanistic – where the mathematical relationship between the input and the output is identified. However, it is only considered representative of the filter if it both explains the data well and provides a physical description of the system under study.

Examination of table 5.1 demonstrates that a deterministic model can not be used for rainfall – runoff modelling in the Trannon catchment due to a lack of parameterisation and parameter data sets. It is also clear that data-based mechanistic (DBM) modelling offers significant advantages over pure data-based modelling and as such DBM modelling has been employed in this study.

5.1.2 Data-based mechanistic modelling of rainfall – runoff: model overview

The DBM modelling strategy applied in this study is that described in Young and Beven, (1994). The model is available as a MATLAB toolbox called 'Captain', from within which the medium term data set for the Trannon has been generated.

The transfer function model used is formulated as follows:

$$y(k) = \frac{B(z^{-1})}{A(z^{-1})} u_e(k - \delta)$$

(equation 5.1)

Where:

k denotes the value of the associated variable at the k^{th} sampling instant

$y(k)$ is the measured stage

Model Type	Advantages	Disadvantages
Deterministic	• able to represent the system well allowing investigation of the role of individual parameters and hence the internal mechanism of the system • if well parameterised, the model output is a physically-based representation of the system	• requires knowledge of the catchment processes prior to parameter identification – *catchment processes in the Trannon catchment have not undergone detailed investigation therefore important parameters are unknown* • data sets pertaining to each parameter are required – *these data sets are not available for the Trannon catchment*
Data-based	• based entirely on the mathematical relationship of the input and output data sets – therefore no physically-based knowledge of the system is required in model generation. • simple to identify and can provide an output which validates well.	• the output is 'black box' in that it bears no relation to the physical processes inherent in the system –*the certainty with which the algorithms represent the filter can not be determined, therefore the output will not respond to changes within the filter.* • a good validation does not mean that the system is well represented.
Data-based Mechanistic	• although based on the mathematical relationship between the input and output data sets it is ensured that mathematical relationship describes the physical processes – *a significant improvement over purely data-based methods meaning that the output is not 'black box'.* • good validation normally means that the physical processes within the system are well represented.	• Model identification is significantly more difficult than pure data-based modelling.

Table 5.1 Summary of major features of available rainfall – runoff model types and their relative advantages in the context of this project.

u_e is the effective rainfall (the rainfall input which has been adjusted non-linearly to account for antecedent soil moisture conditions – see equation 5.3)
δ is the rainfall – runoff delay (lag time)
$A(z^{-1})$ and $B(z^{-1})$ are polynomials in the backward shift operator z^{-1} (which can be thought of simply as a lag operator) [i.e. $z^{-i}\, y(k) = y(k-i)$] which relate rainfall to runoff linearly

A and B are defined by the Captain Toolbox during model structure identification according to the linear system model (see Lai, 1979) in which parameters (a) and (b) are the impulse response weights which are analogous to regression coefficients:

$$A(z^{-1}) = 1 + a_1 z^{-1} + a_2 z^{-2} + \ldots + a_n z^{-n} \; ; B(z^{-1}) = b_0 + b_1 z^{-1} + b_2 z^{-2} + \ldots + b_m z^{-m}$$

(equation 5.2)

in which n and m are integers which can be thought of as defining the system memory. Under normal conditions n and m are seldom >3. The linear system model is used here instead of the more common regression model due to its ability to deal with temporally transient input – output relationships via use of the backward shift operator.

The rainfall – runoff relationship is non linear because antecedent rainfall conditions affect the subsequent flow behaviour. In this model the system non linearity is replicated via the effective rainfall parameter (u_e), which is obtained as a product of the rainfall and the flow (Young and Beven, 1991). Simply, the flow is considered to respond in such a way that its response, relative to that of the preceding rainfall, can be used as a surrogate for soil moisture: low magnitude flows following high magnitude rainfall would suggest a condition of low antecedent moisture. This response is non linear and commonly takes the form of a power function or exponential relationship which is identified by the Captain software.

The above model is extremely simple since it is defined by 3 + n + m parameters. It is data-based since the parameters are estimated directly from the available data. It is also mechanistic in the sense that some of the parameters (i.e. effective rainfall) have a physical interpretation in relation to the physical nature of the system: A and B can be thought of as determining the hydrograph shape and u_e as determining the hydrograph magnitude.

5.1.3 Model definition

850 days of continuous total daily rainfall data collected at the Caersws station and corresponding daily mean stage (DMS) data collected at Dollyd, covering the period April 1994 to September 1996, were used to define the model parameters. Captain toolbox identified the optimum model thus:

- A and B are 1st order polynomials: A = 1.000, -0.7453; B = 0.0115
- u_e is defined as:

$$u_e = u_t \times 1.14 \times \left(1 - \exp\left(\frac{y_t}{0.19}\right)\right)$$

(equation 5.3)

Where

u_t = daily total rainfall at time *(t)*

y_t = daily mean stage at time *(t)*

- $R^2 = 0.798$ i.e. the model explains 79.8% of the data
- Young information criterion (YIC) = -8.6787: a high value for YIC is returned where the model has been over parameterised – clearly this is not a problem with this 1st order polynomial model.

The model output (figure 5.1.) matches the hydrograph shape well suggesting that the polynomials of the A and B parameters are acceptable. It can also be seen that the model replicates the majority of high magnitude events well although it

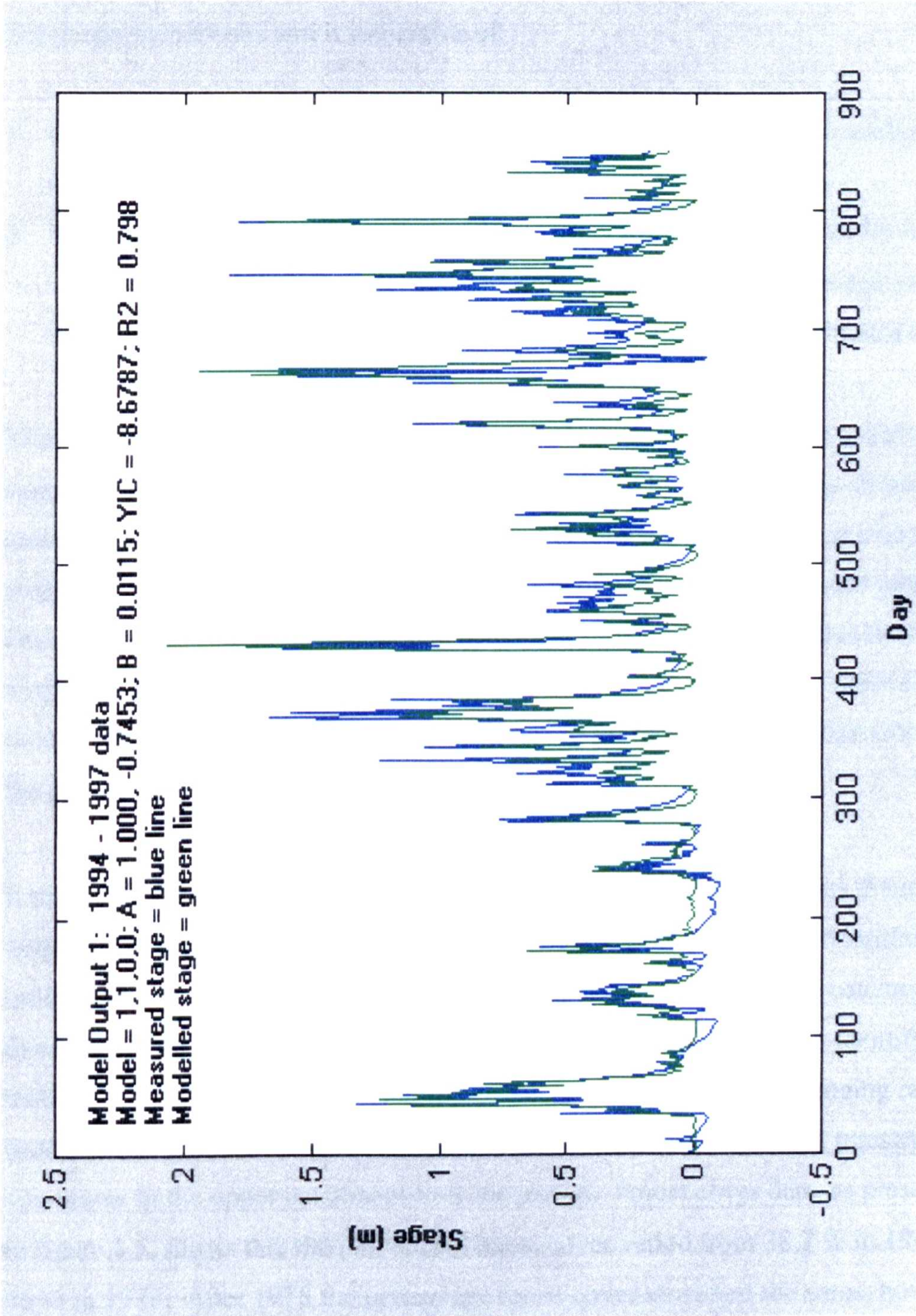


Figure 5.1 Comparative plot of measured and modelled daily mean stage 1994 - 1996

significantly underestimates events at 370 and 750 days whilst overestimating events at 440 and 680 days. This heteroscedasticity, which is a result of high residuals in the non linearity relationship at high stage values, is to be expected when modelling high magnitude events and is indicative of:

1. use of daily total and daily mean data rather than data sets of higher temporal resolution.
2. the use of a rainfall record collected 10km away from the stage recorder for which the flow is being modelled. The spatial variability of high magnitude rainfall events would appear to be greater than that of lower magnitude events.

The fact that the model both under and over estimates high magnitude events suggests that high resolution temporal analysis of unit hydrograph type should be undertaken with caution. However the output is robust enough for time averaged analysis, in which the over and under estimation of peaks will cancel each other out. Due to the data limitations mentioned above the above model represents the best output possible for the Trannon and has therefore been applied to the entire Dollyd rainfall record, creating a daily mean stage record for the Trannon at Caersws for the period 1969 – 2000.

It should be noted that the modelling strategy employed here represents the input - output filter as a temporally static entity as defined by the 1994 – 1996 rainfall – runoff data. The 1969 - 1999 output is therefore representative of the catchment dynamics during 1994 – 1996: the period from which the data used to identify the model were collected. Variation in the output is solely a result of changing rainfall patterns between 1969 and 2000, with no account taken the impact of forestry operations in the upper catchment over the period. Forest cover data, as presented in figure 4.5, shows that the percentage forest cover varied from 38.7 % in 1969 to 56 % in 1978. After 1978 the percentage forest cover remained the same, however felling and replanting of some harvested forest parcels was undertaken. The planting data for the period of the modelled stage record is presented in table 5.2 below.

Table 5.2 Total forest area planted 1969-1999

Year	Total area planted (ha)	Type of planting
1969	35.4	from moorland
1973	72.1	from moorland
1975	73.4	from moorland
1976	7.1	from moorland
1977	7.3	from moorland
1978	7.0	from moorland
1979	7.8	replanting of harvested area
1984	24.7	replanting of harvested area
1985	37.5	replanting of harvested area
1992	58.3	replanting of harvested area
1994	45.0	replanting of harvested area
TOTAL	**375.6**	

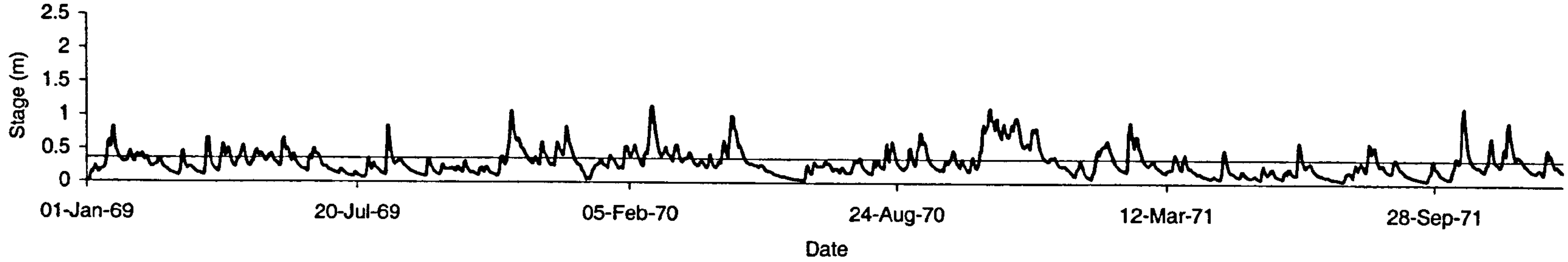
In total 375.6 hectares (equivalent to approximately 35 % of the upper catchment area) was either planted for the first time or replanted over the 30 year period of the modelled record. The periods 1969-1975 (180.9 ha), 1984-1985 (62.2 ha) and 1992-1994 (103.3 ha) contain the highest levels of planting and hence have the greatest potential to have increased peak discharges in the Trannon.

Based on the above data and the findings of Robinson (1980) previously discussed, one might expect peak discharges in the 1994-1996 measured record, used to defined the model, to be slightly elevated as a result of the planting during 1992 and 1994. Consequently, it is feasible that the modelled output may overestimate peak discharges in periods where little planting has occurred (i.e. the late 1970s and early 1980s). However, the amount of planting between 1992 and 1994 is relatively small being only 9.5 % of the total upper catchment area and consequently its impact on peak discharge is likely to be substantially less than Robinson observed in his 100 % planted catchment. Moreover, the measured data used in the model definition were recorded at the downstream end of the Trannon, more than 10 km downstream of the forested upland catchment. As a result, hydrologic inputs form non-forested tributaries are likely to have buffered out much of the magnitude of any enhanced peak discharges by the time they reached the stage recorder. To summarise, the relative small percentage of the upper catchment undergoing planting through the modelled record would suggest any enhancement of peak discharges in the modelled record for the lowland channel would be minor, especially as tributary inputs from non-forested catchments have the potential to buffer out some, if not much, of the increased flood peaks.

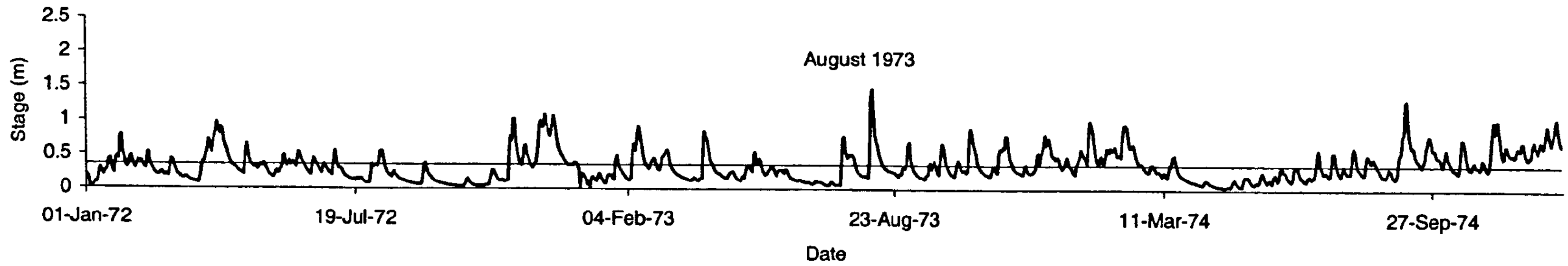
5.1.4 The model output

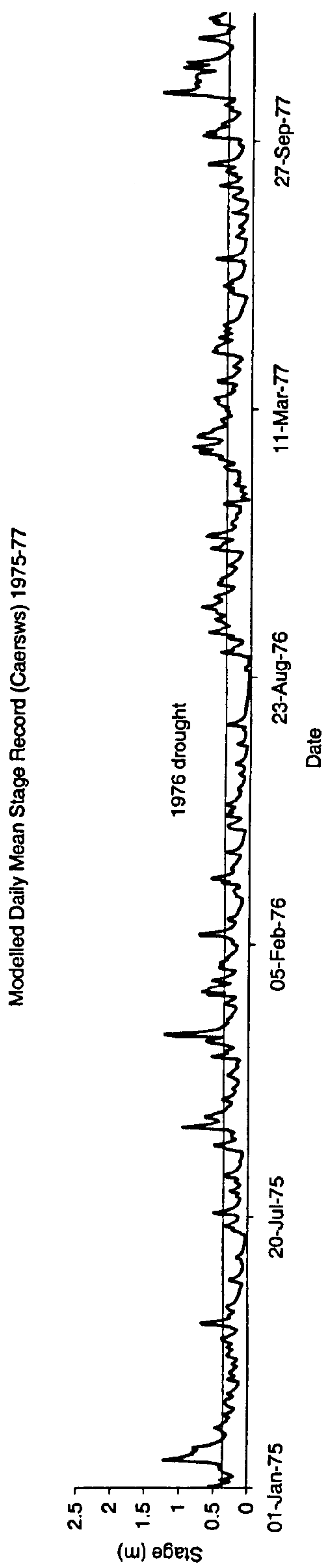
The complete modelled stage record covering 1969 – 2000 is presented in figure 5.2. The mean stage for the 30 year record (0.36 m) is shown as a single line. Initial inspection of the data reveal that seasonality is evident in the output, with the majority of high magnitude events occurring throughout the winter months as compound flood peaks. Individual summer storms can also be identified in the output; for example the flood peak of the 10th August 1973 which is noted in Newson (1980) as being important in elevating bed load yields in the Tanllwyth catchment. The impact of the 1976 drought can also be clearly seen with exceptionally low flows during the period June – August 1976. Indeed the lowest flow recorded (0.00 m) occurs in this period. These features provide evidence that the impact of both short and longer-term changes in rainfall pattern are represented in the model output. More generally, the data suggest an increase in the magnitude of flood peaks after 1993, in particular the peaks occurring on the 28th December 1994 and the 22nd December 1995 exceed 1.5 m stage for the first time. These high magnitude peaks are subsequently a common feature of the measured stage from 1996 – 2000. Summary hydrograph data are presented in table 5.3.

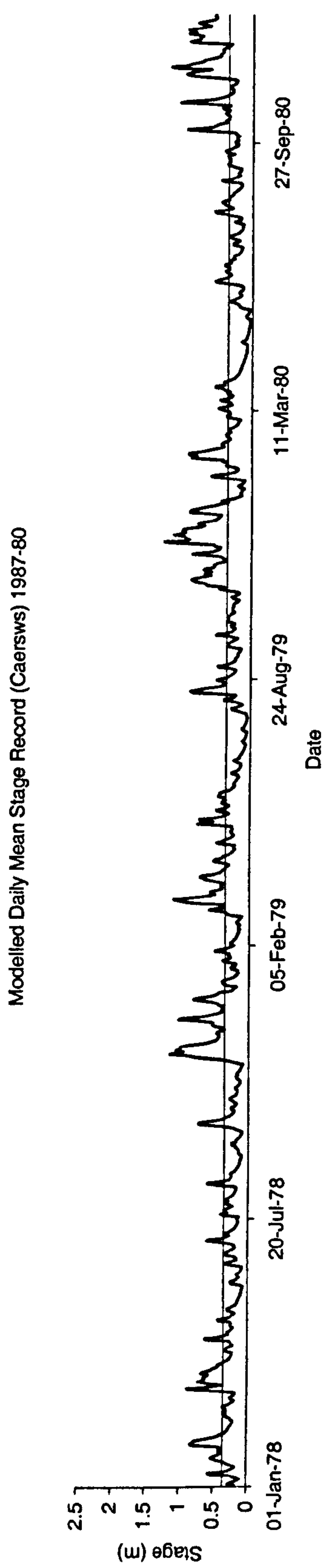
Modelled Daily Mean Stage Record (Caersws) 1969-71

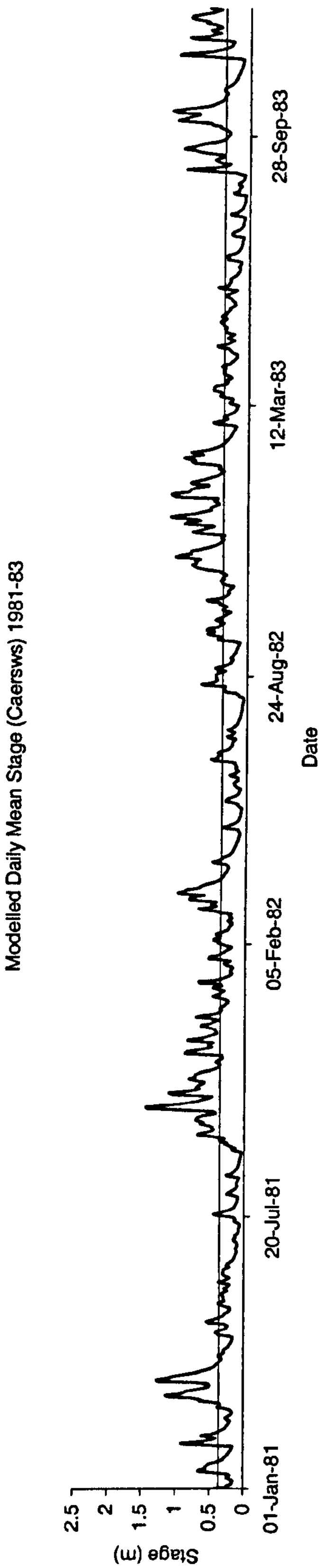


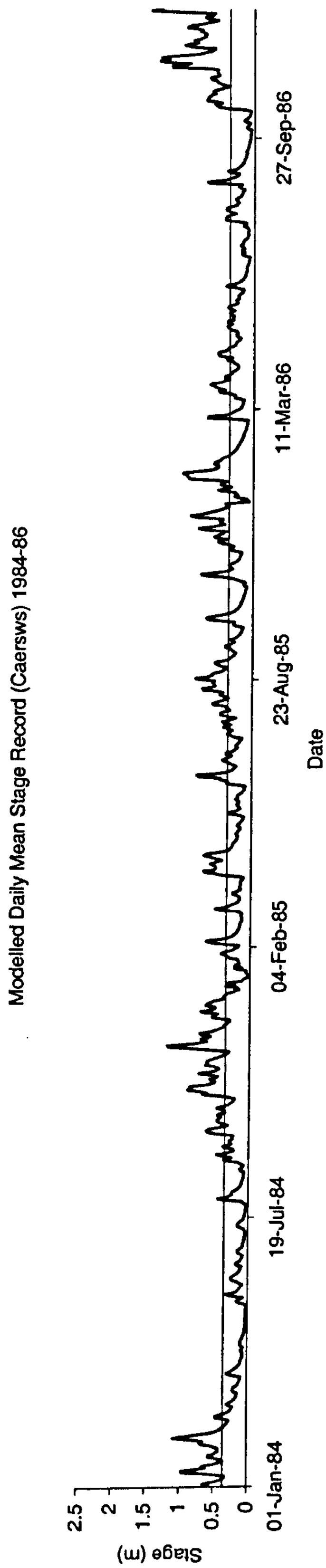
Modelled Daily Mean Stage Record (Caersws) 1972-74



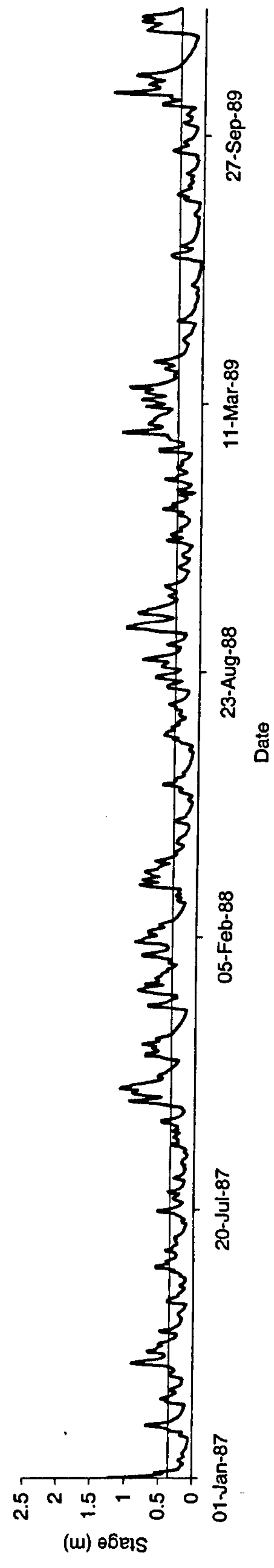

Modelled Daily Mean Stage Record (Caersws) 1975-77
1976 drought
Stage (m)
2.5
2
1.5
1
0.5
0
01-Jan-75
20-Jul-75
05-Feb-76
23-Aug-76
11-Mar-77
27-Sep-77
Date


Modelled Daily Mean Stage Record (Caersws) 1987-80
Stage (m)
2.5
2
1.5
1
0.5
0
01-Jan-78
20-Jul-78
05-Feb-79
24-Aug-79
11-Mar-80
27-Sep-80
Date

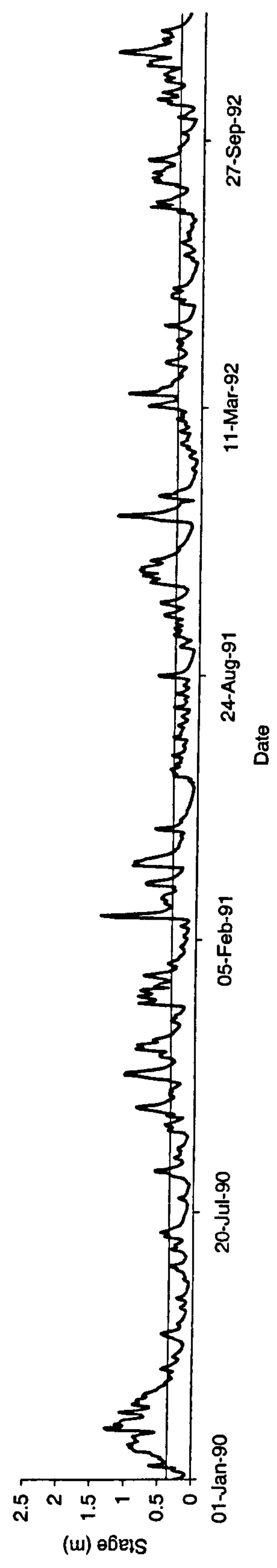

Modelled Daily Mean Stage (Caersws) 1981-83
Stage (m)
2.5
2
1.5
1
0.5
0
01-Jan-81
20-Jul-81
05-Feb-82
24-Aug-82
12-Mar-83
28-Sep-83
Date


Modelled Daily Mean Stage Record (Caersws) 1984-86
Stage (m)
2.5
2
1.5
1
0.5
0
01-Jan-84
19-Jul-84
04-Feb-85
23-Aug-85
11-Mar-86
27-Sep-86
Date

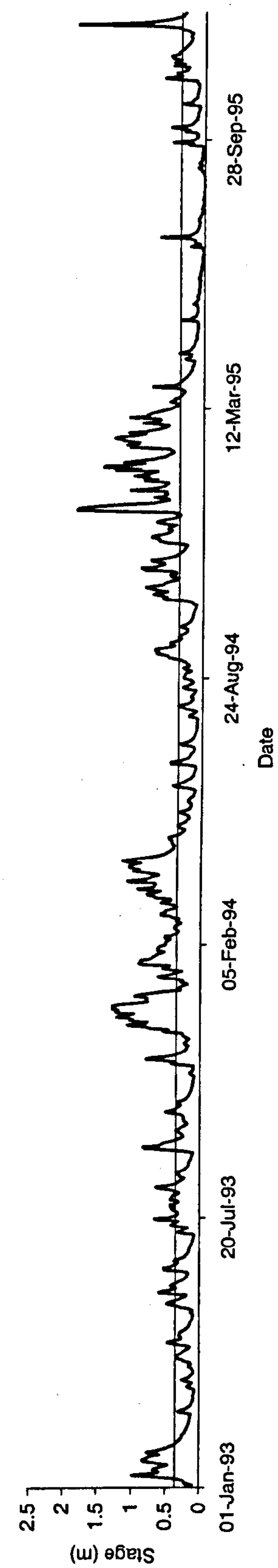
Modelled Daily Stage Record (Caersws) 1987-89



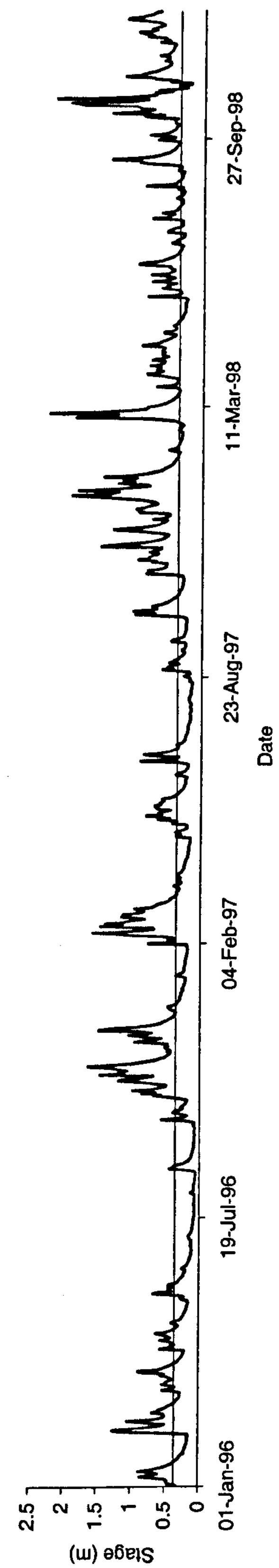
Modelled Mean Daily Stage Record (Caersws) 1990-92



Modelled Daily Mean Stage Record (Caersws) 1993-95



Measured Daily Mean Stage (Caersws) 1996-98



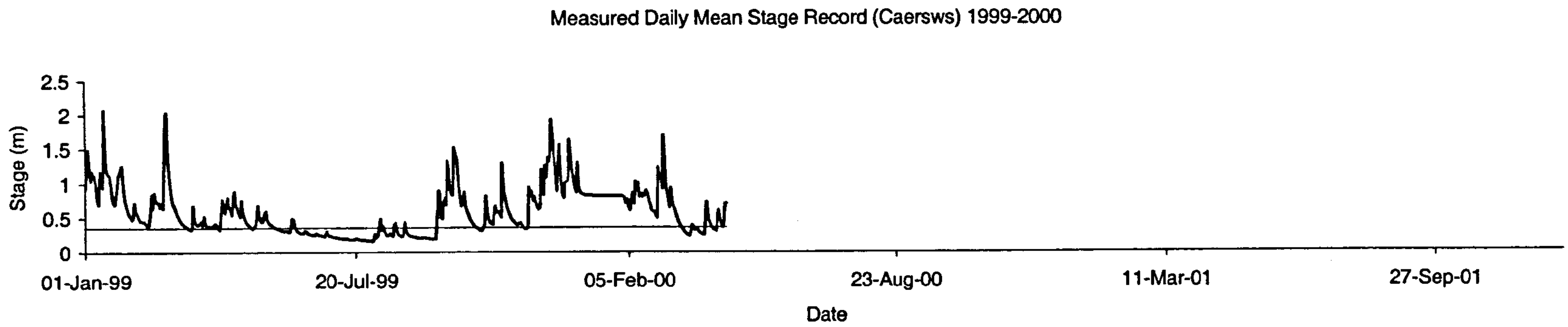


Figure 5.2 this and preceeding five pages: Modelled daily mean stage record of the Afon Trannon at Caersws 1969-1999
The 30 year mean DMS is shown on all hydrographs as a single solid line.

Table 5.3 Summary hydrograph data for the modelled Afon Trannon daily mean stage record.

	Jan 1969 to Dec 1992	**Jan 1993 to Dec 1995**	**Jan 1996 to Feb 2000**
Maximum stage	1.47 m	1.85 m	2.23 m
Minimum stage	0.00 m	0.02 m	0.06 m
Number of floods > 1.5 m	0	2	9
Mean stage	0.33 m	0.35 m	0.49 m

5.2 Determination of flow thresholds

The impact of flood magnitude – frequency on bed load transport rates and channel change is presented in figure 5.3:

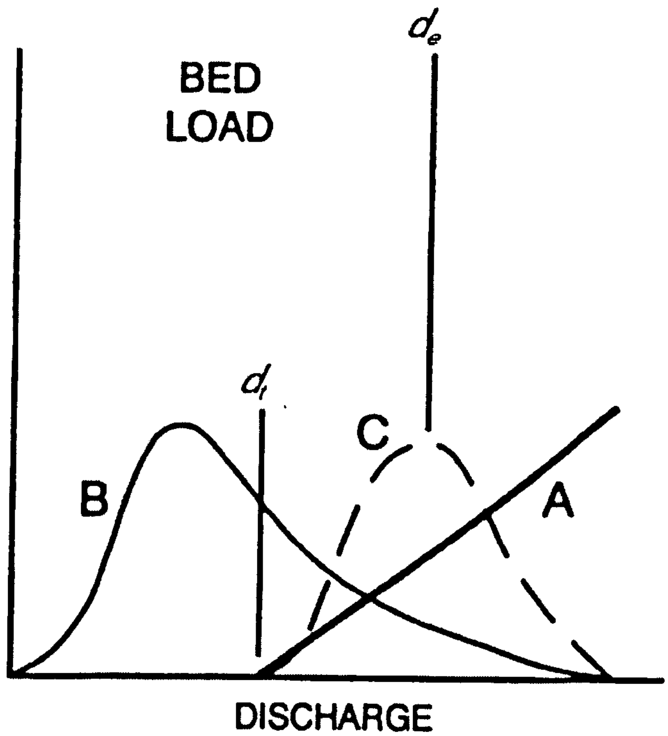


Figure 5.3 The relationship between flood magnitude, frequency and bed load transport rates. A is the bed load transport rate, B is the frequency of occurrence, C is the product of magnitude and frequency, d_e is the stage at which the most effective (or dominant) discharge occurs and d_t is the stage at which the entrainment threshold occurs. After Knighton (1998).

From figure 5.3 it can be deduced that d_e occurs somewhere between the flow magnitude found at the entrainment threshold and high magnitude / low frequency flows. Identifying d_e in the field necessarily requires sediment transport data sets and corresponding flow or stage records for a large number and range of flood events. Such data are unavailable for use in this work. Field identification of the entrainment threshold (d_t) is however much easier, requiring only simple tracing or trapping techniques and corresponding flow or stage records. d_t also indicates the point at which flow becomes responsible for geomorphologic work, therefore examination of the medium-term frequency with which d_t is exceeded can indicate the frequency with which morphological work was achieved, whilst examination of the maximum flood magnitudes can provide an indicator of the magnitude of that work.

5.2.1 Identifying d_t: Methods

Two methods were employed in an attempt to define the DMS threshold above which bed load motion occurred in the destabilised Trefeglwys reaches: small bed load traps and enhanced magnetic susceptibility tracing (Arkell, 1983; Arkell *et al.*, 1985; Leeks *et al.*, 1988).

5.2.1.1 Bed load trapping

Three bed load traps (Tr1 – Tr3) were installed between 10 m and 80 m upstream of the BH reach. The traps used were plastic crates measuring 0.58 m × 0.44 m × 0.52 m which were sunk into the bed and weighted with bricks so that the crate rim was flush with the bed surface. The traps were orientated so that the long axis was orthogonal to flow. Due to the variable nature of channel bed form the traps were located in different channel positions ensuring they remained submerged, even during low flows:

- Tr 1 - on left margin of small unit bar approximately 2 m from bank edge
- Tr 2 – proximal to deepest point on meander bend
- Tr 3 – on margin of unit bar associated with large meander bend

Traps were emptied by hand at approximate intervals of between 2 and 8 weeks. Where possible the traps were emptied after a single flood event, but frequently this was not achieved. Trapped material was air dried and sieved to 0.5 Φ fractions. The fractions were weighed and the b-axis of the largest pebble trapped for each sample was measured and recorded.

5.2.1.2 Enhanced magnetic susceptibility

345.5 kg of bed load was bulk sampled along a transect 7 m long at Pont y Brynllwyn, Trefeglwys. The transect was separated into 1 m units and the bed armour layer (assumed equal to the depth of the coarsest bed particle) was removed separately, from each unit. This meant that the bed was sampled to a depth of between 15 and 20 cm. The sampled material was air dried and sieved at 0.5 Φ mesh intervals. Subsequently, all bed load was heated to 900 °C in a muffle furnace to enhance its magnetic susceptibility according to the methodology of Oldfield et al. (1981). Commonly, susceptibility of individual pebbles was increased from approximately 30χ to 180χ and pebble colour changed from grey to pink. The material was replaced in the channel in its original position ensuring a close match between the tracer and bed load size distribution.

A Bartington M.S.1 coil-type magnetic susceptibility sensor was used to monitor tracer material in the study reach. The susceptibility meter responds to the magnetic flux around a coil sensor which is disrupted when magnetised material protrudes into the space encompassed by the coil. This type of sensor can trace surface particles only. A digital output is provided which, if required, can be used to provide a semi-quantitative estimate of the volume of magnetized tracer beneath the coil (Arkell *et al.*, 1985).

Parallel section lines were established at 1 m intervals downstream of the injection line. Bed magnetic susceptibility readings were taken every 0.5 m across the channel width at the injection line and at each section line both at the time of injection and during the subsequent 2 re-surveys. Where bed susceptibility was higher than the background level a 'hit' was recorded, however extremely high susceptibility values (>1000 χ) suggested the presence of iron or steel objects in the bed material and these were ignored.

5.2.1.3 Accompanying stage record

Throughout the trapping and tracing experiments stage was recorded 60 m upstream of the start of the BH study reach, at the bed load trap location and 120 m downstream of the magnetic susceptibility trace. Stage was recorded using a pressure transducer type apparatus and automatic logger. A Druck PDCR 1830 pressure transducer, with a 75 mbar operating range, and an accuracy of ± 0.1 % was housed in a 2 m high portable stilling well of 150 mm diameter. An integral stage board was housed on the stilling well. Transducer output (mV) was logged every 1 minute by a Grant Squirrel 1000 logger. Readings were time averaged every 15 minutes and the time averaged reading was stored. The transducer excitation voltage, provided by a 9V battery, was also logged concurrently.

The transducer was calibrated at 0.2 m depth intervals to a depth of 2.4 m in a swimming pool. Because excitation voltage in the field was variable (due to battery drainage) the calibration procedure was repeated for every 0.2 V from 9.5 V to 2 V. Voltage was supplied by a Thurlby PL310 power pack with an output accuracy of +/- 0.01 V. Transducer output was monitored on the Squirrel logger and a variable voltage conversion formula was derived.

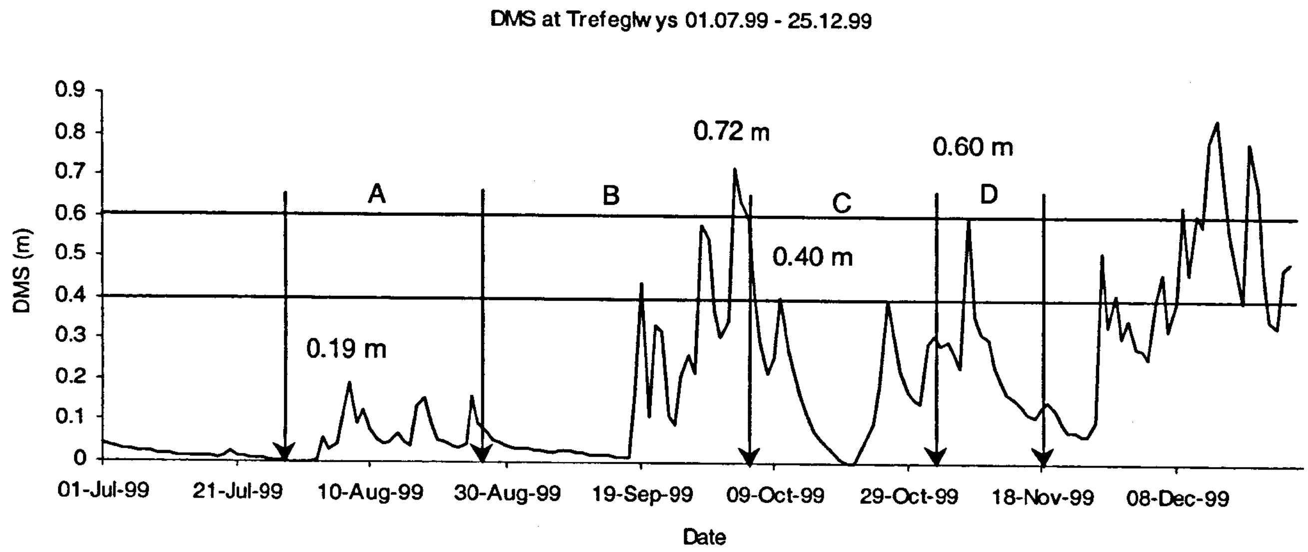
5.2.2 Identifying d_i: bed load trap results

The bed load trap results are presented in table 5.4 with the accompanying stage record from the Trefeglwys stage recorder presented in figure 5.4. The results for each trapping period are summarised below:

Period A:

Only trap 2 was installed in the bed. 4.4 kg of material was collected and the maximum daily mean stage attained in the period was very low (0.18 m). In period A trap 2 was located next to a cattle drinking area, therefore the small trap yield is thought to be due to cattle kicking material into the trap and the result is discounted.

Figure 5.4 Daily mean stage (DMS) record covering the bed load trap data collection period. The four trapping periods are labelled A-D. Also shown is the maximum DMS value attained in each trapping period.



Period B:

All traps were installed and functioning. All traps collected material with the largest quantity collected in trap 1 (31.7 kg). This is unsurprising given the central location in the channel of this trap. Particles up to a D_{max} of 102 mm were collected inferring that even the larger bed fractions were in transport. The maximum daily mean stage attained in period B was 0.72 m. It is considered that in this period the entrainment threshold had been exceeded, and hence 0.72 m daily mean stage was greater than the entrainment threshold.

Period C:

All traps remained empty in period C meaning that the entrainment threshold was not exceeded. The maximum daily mean stage attained was 0.40 m, therefore the entrainment threshold can be said to exist at a daily mean stage greater than 0.40 m and less than 0.72 m.

Period D:

Traps 1 and 2 were functioning. Trap 3 had been damaged. Both traps 1 and 2 collected bed material (47.6 kg and 10.26 kg respectively) demonstrating that the entrainment threshold was exceeded in this period. The D_{max} collected was 98 mm, again suggesting that the coarse bed fractions were experiencing some transport. The maximum daily mean stage recorded in the period was 0.60 m – less than that in period B. Therefore, the entrainment threshold can be more closely identified as existing between 0.40 m and 0.60 m daily mean stage.

Table 5.4 Bed load trap results.

Trapping period	Trap 1		Trap 2		Trap 3	
A	no data		Total mass	4.4 kg	no data	
(29.07.99-05.10.99)			D_{max}	42 mm		
B	Total mass	31.7 kg	Total mass	8.6 kg	Total mass	14.3 kg
(27.08.99-05.10.99)	D_{max}	71 mm	D_{max}	92 mm	D_{max}	102 mm
C	Total mass	Empty	Total mass	Empty	Total mass	Empty
(06.10.99-04.11.99)	D_{max}	Empty	D_{max}	Empty	D_{max}	Empty
D	Total mass	47.6 kg	Total mass	10.3 kg	no data	
(05.11.99-18.11.99)	D_{max}	98 mm	D_{max}	54 mm		

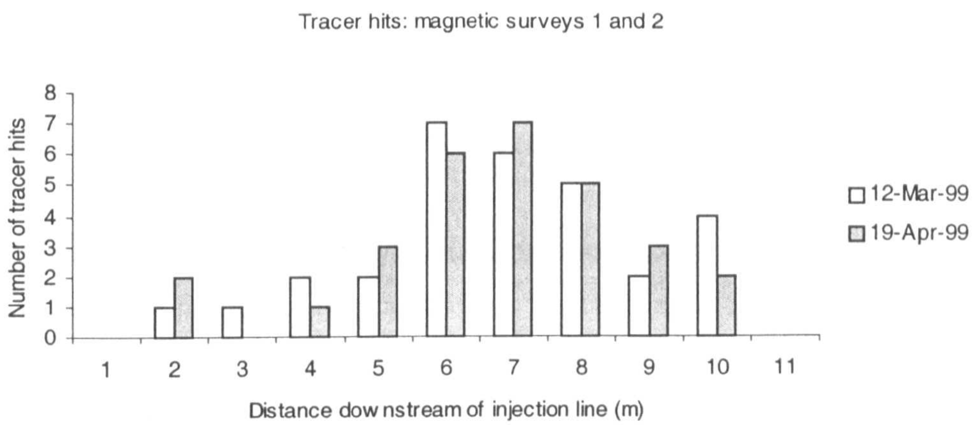
5.2.3 Identifying d_t: magnetic trace results

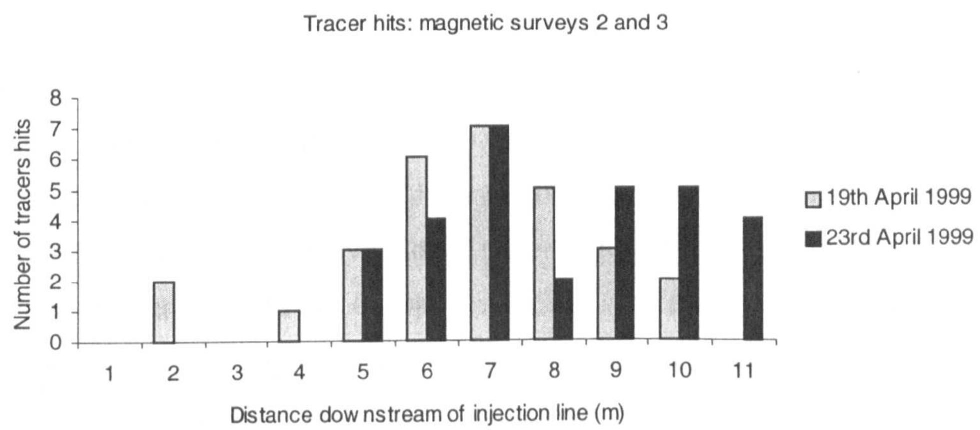
Magnetically enhanced bed load was injected into the channel on 14.02.99 and three subsequent bed surveys were undertaken: Survey 1 – 12.03.99; Survey 2 – 19.04.99 and Survey 3 – 23.04.99. The results of these surveys for sections 1 – 11 m are presented in figure 5.5 together with the mean difference in the number of tracer hits for each survey compared to survey 1 which provide a simple assessment of similarity (table 5.5).

Table 5.5. Mean difference in number of tracer hits between surveys. The lower the value the greater the similarity in the data distribution. A value of 0 shows and identical distribution.

Tracer distributions being compared	Survey 1	Survey 2
Survey 2	0.81	0.00
Survey 3	1.81	1.54

Figure 5.5 Number of tracer hits recorded between magnetic bed load surveys 1 and 2 and surveys 2 and 3.

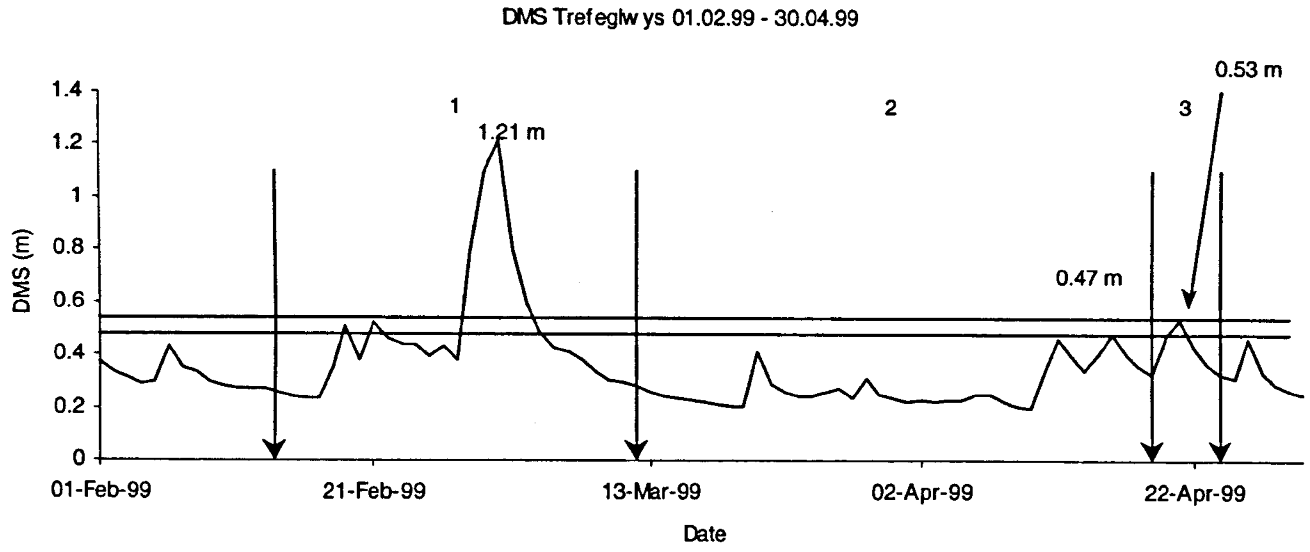




The downstream pattern of tracer hits for surveys 1 and 2 is extremely similar with mean difference value of 0.81 (a mean difference value of 0.0 shows that the distribution is identical). The downstream distribution of tracer hits approximates normal in both surveys and tracer material is first detected 2 m downstream of the injection line. Clearly, there has been substantial bed load transport between tracer injection and survey 1, however the variance in tracer distributions between surveys 1 and 2 is most likely explained by variation in the sampling positions between the survey dates: it is almost impossible to relocate each bed sample position exactly between surveys. There is little evidence, therefore, that tracer motion has occurred between surveys 1 and 2 suggesting that d_t was not exceeded in this period.

Tracer distributions for surveys 2 and 3 are less similar (mean difference = 1.54). Whilst tracer material was located at 2 m from the injection line in survey 2, no tracer material is located until 5 m in survey 3. The approximately normal tracer distribution of survey 2 is replaced by a right skewed distribution for survey 3. Also, tracer material is found further downstream at 11m from the injection line in survey 3 but it is absent in survey 2. As such, there is strong evidence that tracer material underwent transport in the period between surveys 2 and 3, and that d_t was therefore exceeded. Placing these tracer results in a hydrologic context (figure 5.6) locates d_t to within a small range of DMS. The maximum DMS recorded between surveys 1 and 2 is 0.47 mm and the tracer results strongly suggest little or no tracer movement. Between surveys 3 and 4 tracer results strongly suggest downstream transport of tracers and this transport coincides with a maximum recorded DMS of 0.53 m. Consequently, the data identify d_t as occurring within the range 0.47 – 0.53 m DMS; a figure which also falls within the range identified by the bed load trapping experiments (0.40 m – 0.60 m). The entrainment threshold is assigned the maximum value of 0.53 m daily mean stage as recorded at the Trefeglwys stage recorder.

Figure 5.6 Daily mean stage (DMS) record covering the period of the magnetic tracer experiments. Tracer injection occurred on 14.02.99 and is indicated by the firest arrow. Inter-survey periods 1-3 are identified together with the maximum DMS attained in each period.



5.3 DMS > d_t: medium-term frequencies

5.3.1 The relationship between DMS at Trefeglwys and Caersws

Examination of the frequency with which flow exceeds d_t throughout the 1969 – 2000 modelled DMS record, allows conclusions to be drawn about the temporal patterns of channel change within a hydrologic context. The identification of d_t, presented previously, has been in relation to a stage record gathered at Trefeglwys, some 5 km upstream of the Caersws stage recorder for which the medium-term DMS record has been generated. Therefore the instantaneous relationship between stage at Trefeglwys and Caersws must be established and applied to d_t in order to quantify the frequency with which d_t was exceeded at Trefeglwys over the period 1969 – 2000.

The relationship between DMS at Trefeglwys and Caersws are presented in figure 5.7. The data is fitted well ($R^2 = 0.85$) by the linear formula $y = 1.6814x - 0.0019$ in which y is DMS at Caersws and x is DMS at Trefeglwys. Therefore, the entrainment threshold of 0.53 m DMS at Trefeglwys is represented by DMS at Caersws of 0.89 m. In subsequent analysis d_t will be assumed to occur at 0.89 m providing *minimum* frequency estimates.

5.3.2 DMS > d_t: annual frequencies

Figures 5.8 and 5.9 present the frequency (f) with which DMS exceeded d_t (0.89 m), for the period 1969 – 1999. Between 1969 and 1989 the plot displays relative stability (compared to 1989 – 1999) in so much as the frequency varies within a range of 23 days, has a mean of 10.23 days and a standard deviation of 6.06 days. In contrast, between 1989 and 1999 the plot is unstable – displaying an increasing trend. The annual frequency with which d_t is exceeded has a range of 57 days, a mean of 24.27 days and a standard deviation of 16.95 days. Examination of just the 1989 – 1999 data (figure 5.9) reveals an exponential rise in the annual frequency

Figure 5.7 The relationship between daily mean stage at the Trefeglwys and Caersws logging stations.

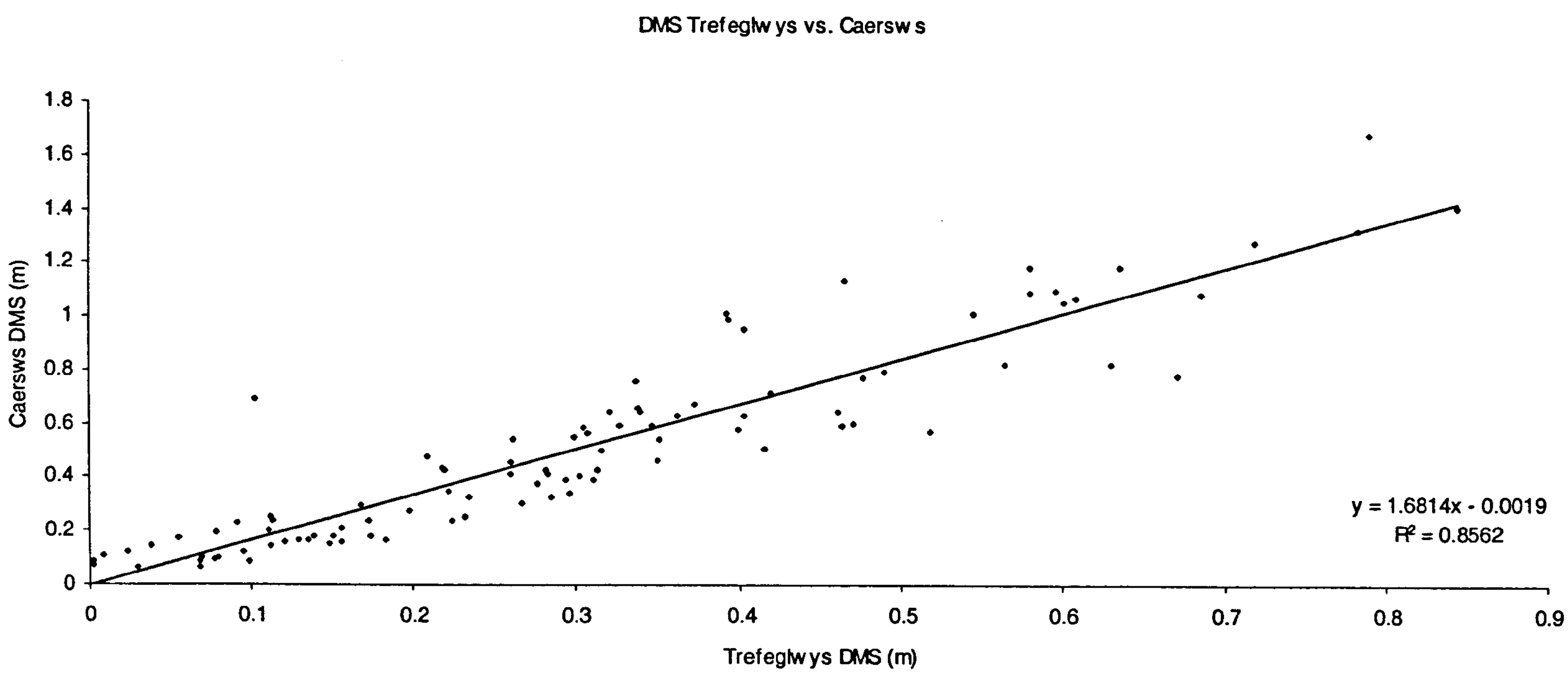


Figure 5.8 The annual frequency with which d_t (0.89 m) was exceeded 1969-1999.

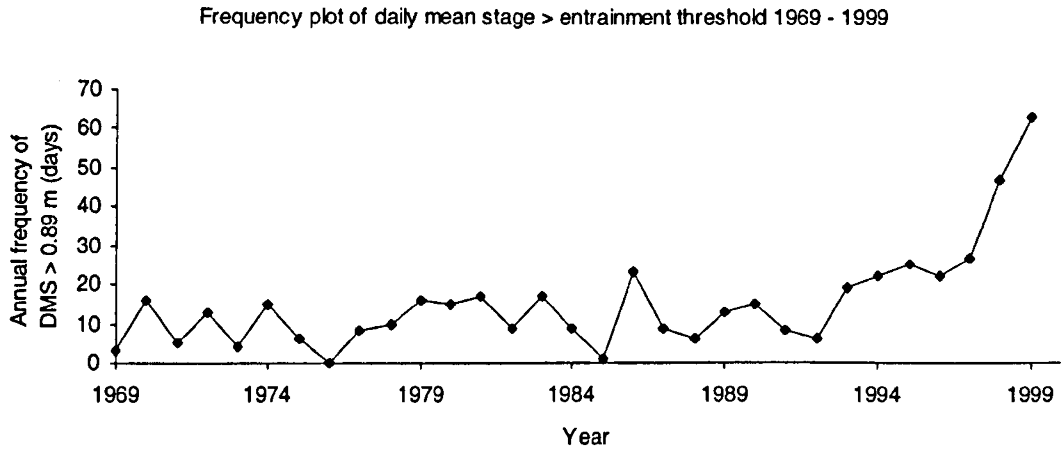


Figure 5.9 The annual frequency with which d_t is exceeded 1989-1999. The increasing frequency is apparent in both the modelled (shaded points) and the measured (unshaded points) data.

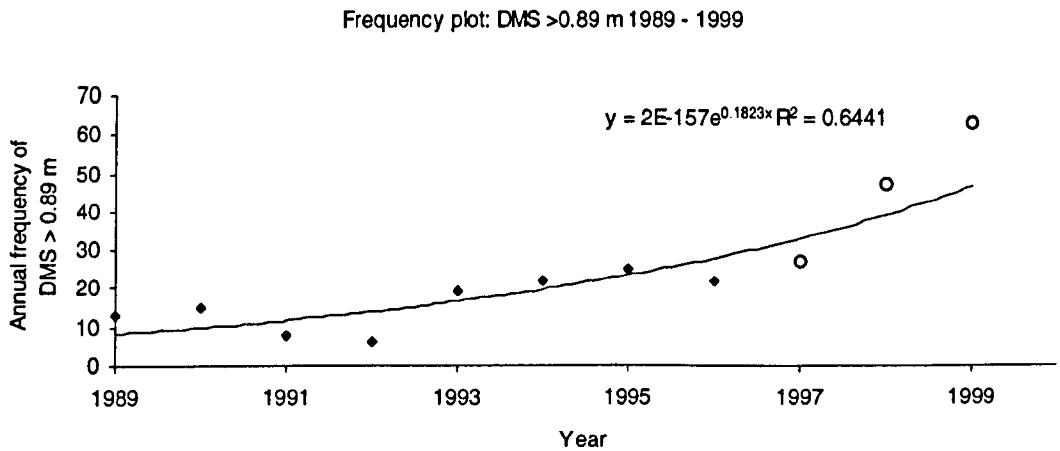
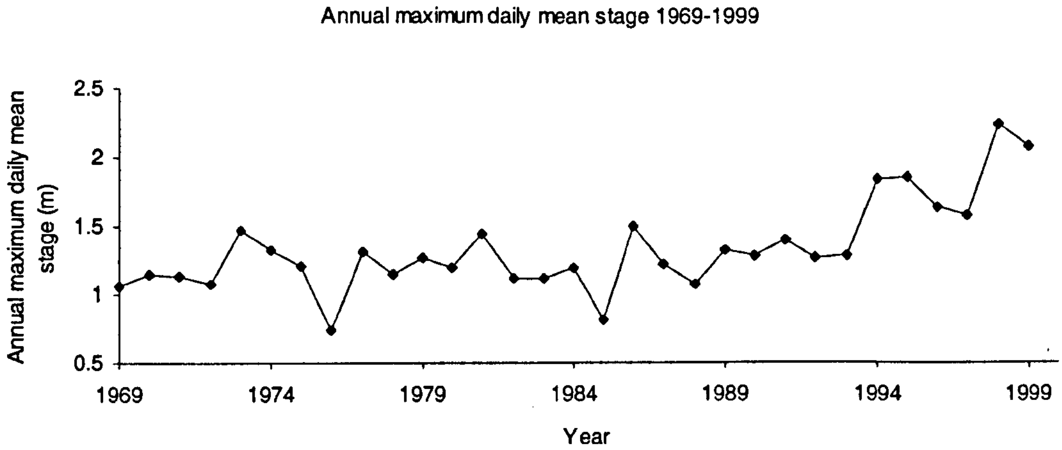


Figure 5.10 Maximum annual daily mean stage for the period 1969-1999.



with which DMS of 0.89 m is exceeded. This exponential rise in stage exists in both the modelled and measured stage data, strongly suggesting that it is not a result of model deficiencies but a real effect of rainfall and catchment antecedent conditions.

5.3.3 Medium-term patterns of flood magnitude

The identified trend of increasing frequency of d_t between 1989 and 1999 is also expressed in the maximum annual DMS (DMS_{max}) (figure 5.10). Again DMS_{max} between 1969 – 1989 can be characterised as relatively stable. DMS_{max} ranges between a minimum of 0.73 m (the drought year of 1976) and 1.50 m with a mean of 1.18 m and a standard deviation of 0.18 m. Between 1989 and 1999 the trend is one of increasing DMS_{max}, and the trend can be poorly described by the linear relationship ($y = 0.0706x-139.23$; $R^2 = 0.68$). DMS_{max} ranges between 1.26 and 2.23 m, has a mean of 1.61 m and a significantly higher standard deviation of 0.34 m.

5.3.4 Annual patterns of competent flow magnitude and frequency: summary

Annual patterns of total rainfall at Dollyd and annual patterns of flow magnitude and frequency have been compared with their respective annual means for the period 1969 – 1998 (see figure 5.11 – 5.13). The pattern of annual frequency with which d_t varies about the 1969 – 1998 mean and the annual variation of DMS_{max} compared to the mean are similar. Between 1969 and 1993 annual values are commonly below the mean: the annual frequency of d_t exceeds the mean frequency for only 8 of 25 years, of which only 1986 is significantly greater than the mean. Similarly, between 1969 and 1993 DMS_{max} only exceeds the mean for 7 years and never exceeds it by more than 0.2 m. By contrast, between 1994 and 1998 both the frequency with which d_t is exceeded and the magnitude of DMS_{max} are significantly above the mean and display an increasing trend.

The results provide strong evidence that the hydrologic loads of the Trannon can vary greatly – suggesting that temporal variability in flow magnitude and frequency

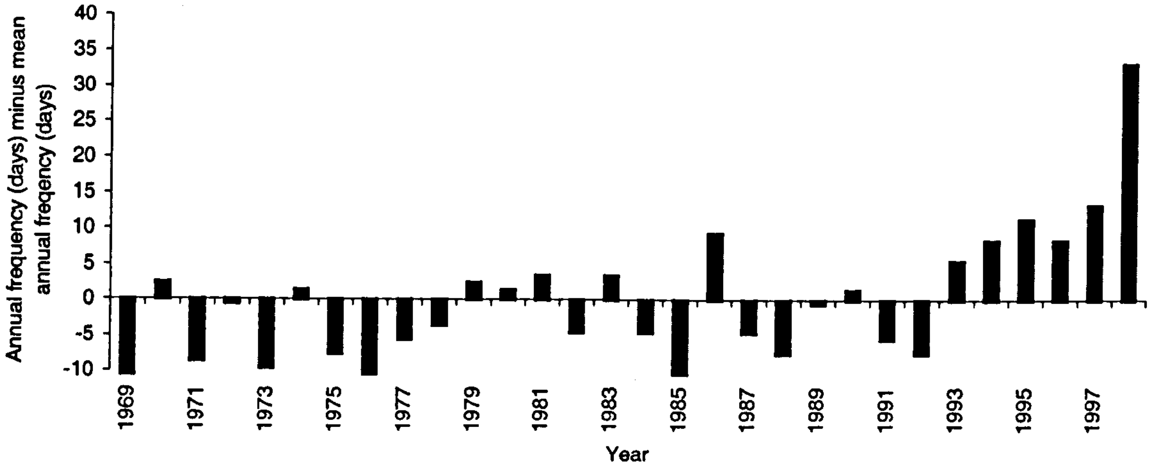


Figure 5.11 Annual frequency minus 1969-1999 mean frequency: DMS > 0.89 m

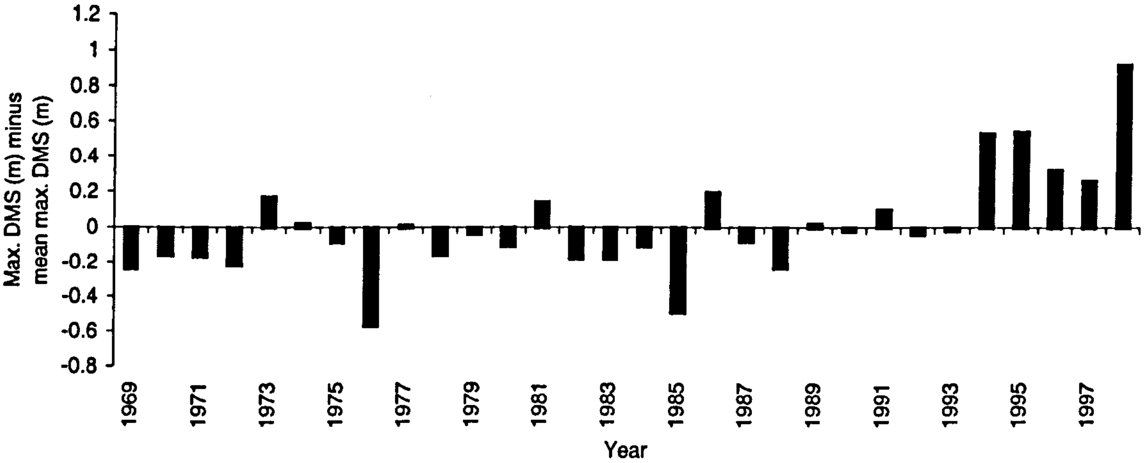


Figure 5.12 Annual maximum DMS minus mean maximum DMS 1969-1999

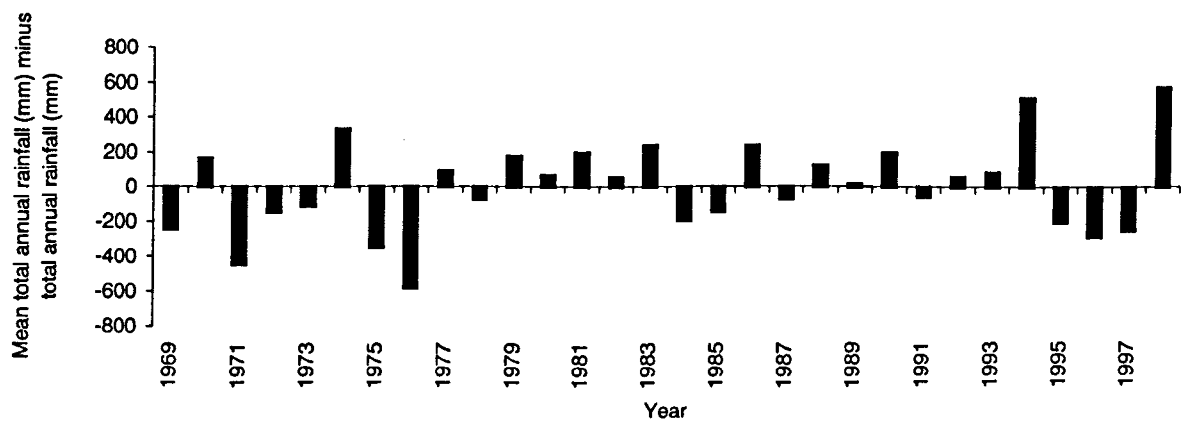


Figure 5.13 Total annual rainfall minus 1969-1998 mean total rainfall.

may play an important part in channel morphology. It can be expected that, if examined, the rates of channel change will be significantly greater for the period 1994 - 1999 than those for the period 1969 – 1993 and that rates of channel change are unlikely to vary greatly between 1969 and 1993 unless in response to increased sediment loads. Decreasing rates of channel change throughout the 1994 - 1999 period would only result from a substantial decrease in recent sediment loads acting to counter the effect of increased magnitude and frequency of competent flows, or as a result of successful channel engineering.

The data also allow implications about the recent temporal patterns of rainfall to be drawn. The increased magnitude and frequency of flows between 1994 and 1998 is not easily predicted from total annual rainfall, where only 1994 and 1998 have total annual rainfall above the mean (+514 mm and +575 mm respectively), with 1995 – 97 well below (-203 mm, -284 mm and –252 mm respectively). This contrasting pattern implies that the increase in high magnitude and frequency flows between 1995 – 97 are a complex result of an increase in the number of high magnitude rainfall events falling onto a catchment with increasing antecedent moisture conditions (indicating the importance of the system non linearity), rather than as a simple response to an increase in the total amount of rain. This would infer that the distribution of rainfall events throughout the year has changed in the last 30 years, with rainfall events becoming more concentrated into shorter periods of time.

6. ASSESSING HISTORIC CHANNEL STABILITY: MEDIUM-TERM PATTERNS OF CHANNEL CHANGE

6.1 Introduction

This chapter presents the methods used in, and the results from, medium-term quantification of channel changes of the Afon Trannon. The data are derived from archive sources, in particular historical aerial photography. Calculation of net rates of change of two dimensional channel parameters is attempted using raw image interpretation, digital photogrammetric correction techniques and a GIS. The procedures allow characterisation of channel stability, at both mid and lowland river channel positions, both prior to and subsequent to upland afforestation.

The aerial photography available for use in this project is summarised in table 6.1. Archive aerial photographic coverage of the Trannon channel is extremely variable, ranging from almost total coverage in 1976, to coverage of only two discrete reaches at Llawr-y-Glyn (680 m) and Trefeglwys (2020 m) in 1948. This limited 1948 coverage has determined the spatial extent to which archive aerial photograph analysis has been undertaken. It also represents the lowland channel setting in the first year of upland catchment afforestation, and hence is able to provide information on characteristics of the lowland Trannon channel uninfluenced by upland catchment forestry.

The Llawr-y-Glyn reach covers 680m of channel between 8.5 and 9.2 km downstream of the Trannon source. Leeks et al. (1988) state that bed load transport is rapid through this reach and that there is little bed load storage. Conversely, significant bed load storage, in the form of gravel barforms, is reported in the 2000 m long Trefeglwys reach. This 2000 m stretch of channel forms the lowland study reach and represents the uppermost storage zone for bed load. It is therefore an important study reach because it represents the location furthest upstream in which channel change resulting from altered bed load supply may be expressed.

Date	Flight I.D.	Flight Alt.	Geometric Fidelity?	Stereo Pairs?	Quality?	Correction?	2D / 3D?	Coverage and comments
1948 (24.03.48)	CPEUK 2531	12000'	NO	NO	VERY POOR	GEO	2D	Approx. 600 m of channel at Llawr-y-Glyn, 2000 m of channel covering Trefeglwys study reach and extending downstream. Images contain production faults in the form of drying patterns.
1963 (22.01.63)	58/RAF/ 5607	12000'	NO	YES	POOR	GEO	2D	4000 m of channel covering Trefeglwys study reach and extending downstream to below Ddranen Ddu. No coverage at Llawr-y-Glyn. Images taken during snow cover.
1976 (28.04.76)	76 046	7900'	YES	YES	GOOD	ORTHO	3D	Approx. 9000 m of channel from Llawr-y-Glwn, extending downstream through Trefeglwys to 1000 m below Ddranen Ddu.
1988 (23.06.88)	88-197	8400'	YES	NO	VERY GOOD	GEO	2D	3500 m of channel covering Trefeglwys study reach and extending to Ddranen Ddu. No coverage at Llawr-y-Glyn.
1995 (07.05.95)	95-541	9600'	YES	YES	VERY GOOD	ORTHO	3D	Approx. 800 m of channel at Llawr-y-Glyn, 4500 m of channel covering Trefeglwys study reach and extending below Ddranen Ddu.

Table 6.1 Summary table of aerial photographic coverage of the Afon Trannon.

The quality of aerial photography available for use in this study is extremely inconsistent, ranging from stereo-pair images with geometric fidelity (1976 and 1995) to non stereo-pair images, taken without a mapping camera and containing no geometric fidelity (1948 and 1963). Where non stereo-pair imagery exists, and photographs do not have geometric fidelity, only two dimensional area measurements can be made. Two dimensional measurements from raw imagery are prone to systematic and non-systematic errors caused by camera orientation, terrain relief, earth curvature, film and scanning distortions (Konecy and Lehmann, 1984; Kraus, 1984, Wang, 1990; Jensen, 1996). These errors can be very large. The use of geo-correction procedures can substantially reduce this error. Geo-correction may be applied to correct for image distortions in the X and Y directions by 'stretching' the image to fit a set of ground control points of known relative position. However, the process does not remove distortion due to land surface change in the Z direction and hence in areas of variable relief measurement from geo-corrected images is extremely unreliable.

Where stereo pair images are available ortho-rectification and subsequent digital terrain model (DTM) construction can be achieved. Ortho-rectification is able to correct for distortions in the X, Y and Z directions which are present in the image. This is a significant advantage over geo-correction and produces a superior quality image, particularly in regions of high topographic relief. Although there are several methods available for ortho-rectification (see Yang, 1997) the collinearity method produces the most reliable solution by incorporating camera orientation, relief displacement and the Earth's curvature in its modelling process (Anon., 1999). The collinearity method, as available in Imagine OrthoBASE version 8.3.1., is applied here to ortho-rectify the raw images. The net result is a digital image for which the inherent distortions have been enormously reduced and for which the relative positions of each pixel are known in the X, Y and Z directions.

In order for ortho-rectification to be undertaken the raw images will ideally have geometric fidelity, fiducial marks (marks which act as a fixed basis of spatial comparison used in rectification procedures) and have been taken by a calibrated mapping camera with known lens and camera distortions. Generally speaking, the more ground control points (a point on the Earth's surface where both image coordinates and map coordinates can be fixed) and tie points (a point on the Earth's surface where image coordinates can be fixed) used during image rectification, the better the result and the lower the root mean square error (RMSE) values associated with the rectified image. The RMSE error can be thought of as an expression of the positional accuracy of any image pixel relative to its real position determined by the mapping grid used in the rectification process. Ground control can be provided by a number of methods including O.S. maps and field survey using differential global positioning systems (DGPS). The accuracy with which the ground control point is placed largely determines the absolute error associated with the ortho-rectification process. The increasing development of GIS means that ortho-images can easily be converted to three dimensional digital terrain models via surface interpolation. The manipulation of ortho-images and DTMs is becoming an increasingly important area of fluvial geomorphology (e.g. Lane et al., 1994; Lane, 2000; Westaway et al., 2000).

6.2 Selected parameters

At each site the following channel morphological parameters have been quantified:

- Bankfull width (as applied by Reinfelds (1997) where bankfull occurs at the break of slope where small increases in stage are accompanied by disproportionately large increases in width (Osterkamp and Hedman, 1982))
- Channel center position (defined as the point halfway between left and right bankfull)

At Trefeglwys the following parameter has also been quantified:

- Sub-aerial gravel bar area

A range of procedures for quantifying these channel parameters from aerial photography have been applied:

1. Two dimensional parameter quantification from raw imagery in PaintShop Pro 5.1 (PSP)
2. Two dimensional parameter quantification from geo-corrected imagery in ERDAS Imagine GIS
3. Two dimensional parameter quantification from ortho-rectified imagery in ERDAS Imagine GIS

Two dimensional parameter quantification from raw imagery within PSP was initially undertaken due to the simplicity and speed with which long distances of channel can be assessed. In this way an initial estimate of relative channel change could be made and particularly active reaches could be highlighted for more detailed examination within a GIS, using the more accurate rectified images which are time consuming and technically demanding to produce. Subsequently, data from the both raw and rectified imagery was compared in order to determine the error inherent in using non rectified imagery for two dimensional analysis and to investigate the possibility of combining the data sets.

Prior to analysis, aerial photograph contact prints were digitised to raw raster images on a UMAX 1220p flat bed scanner. Images were aligned on the scanner bed in one of two ways:

1. Images for use in PSP analysis – all photographs to be scanned were aligned to a common north thus ensuring a homogenous orientation for all raster images. North was derived by drawing a line along a straight stretch of road which appeared on all photography. This 'north' line was then aligned to the scan bed margin.

2. Images for use in ERDAS Imagine analysis - photographs were aligned on the scan bed so that if fiducial marks were present they were within the boundaries of the scanning bed.

All images, independent of analysis type, were scanned at 600 x 600 dots per inch (dpi) so that each pixel represented a square on the image 42 μm $\times$ 42 μm. For a 1:10000 scale photograph (an approximate mean scale of the Trannon imagery) the scanning procedure produced a ground resolution of approximately 0.42 m. This meant that individual pixels in the raster image which were assigned a grayscale value were 0.42 $\times$ 0.42 m in size.

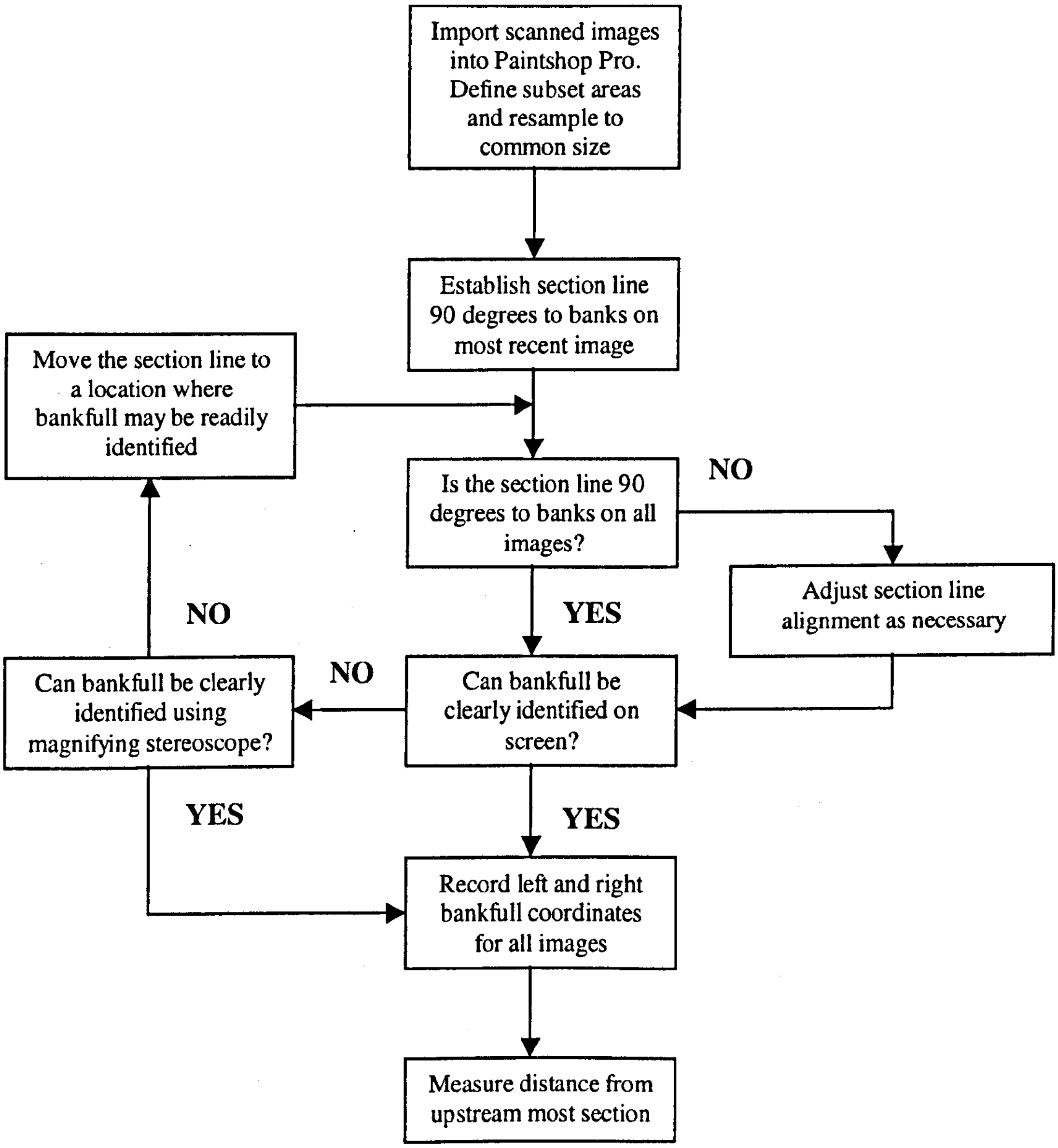
6.3 Parameter quantification within Paintshop Pro (PSP) v5.01 - procedures

The procedures described here are also presented schematically in figure 6.1. For each date the river channel study reaches at Llawr-y-Glyn and Trefeglwys were divided into a number of subsets using the PSP crop tool. The purpose of this was two-fold:

1. To reduce the image file size allowing a faster image rebuild speed
2. To reduce the on screen image size allowing easier navigation about the image

To ensure that the ground area contained within each cropped subset was the same for all image dates the crop tool was dragged from a fixed start feature to a fixed end feature, located at diagonally opposite corners of the crop area. The start and end points were commonly field boundary intersects or the corners of buildings, and were visible on all image dates. The distortions inherent in the raw images and the varying scales of imagery meant that the width and height of subsets varied with image date. Consequently, the cropped areas were resampled and enlarged so that they were of the same size and pixel resolution as the largest scale 1976 imagery. Each image subset then covered the same ground area, at the same scale and resolution for all image dates. Knowledge of the resampled image dimensions and

Figure 6.1 Paintshop Pro procedure for measuring bankfull location.



the original scale of the 1976 imagery and the elevation above mean sea level of the lowland valley, allowed the ground area covered by each pixel to be estimated for each date. This information is provided in table 6.2.

Table 6.2. Image pixel resolutions used in PSP analysis

Image Date	Pixel Resolution (m)
1948	0.50
1963	0.50
1976	0.33
1988	0.35
1995	0.40

Measurements of channel width and lateral change were made using a system of section lines, drawn orthogonal to the channel direction, and onto all images in identical locations (figure 6.1). Pixel coordinates for the start point of each section line and the azimuth of the section (as displayed on screen in PSP) were used to ensure that the section lines were located in the same position for all image dates. Approximately 25 % of section lines had to be rotated about the channel center to ensure that following channel movement the section lines remained orthogonal to the banks and bankfull width changes recorded were not just an artifact of planform change. Maximum rotation was applied to Trefeglwys aerial section number 29 (15°), required as a result of meander migration between 1948 and 1976, but commonly rotations were below 10°.

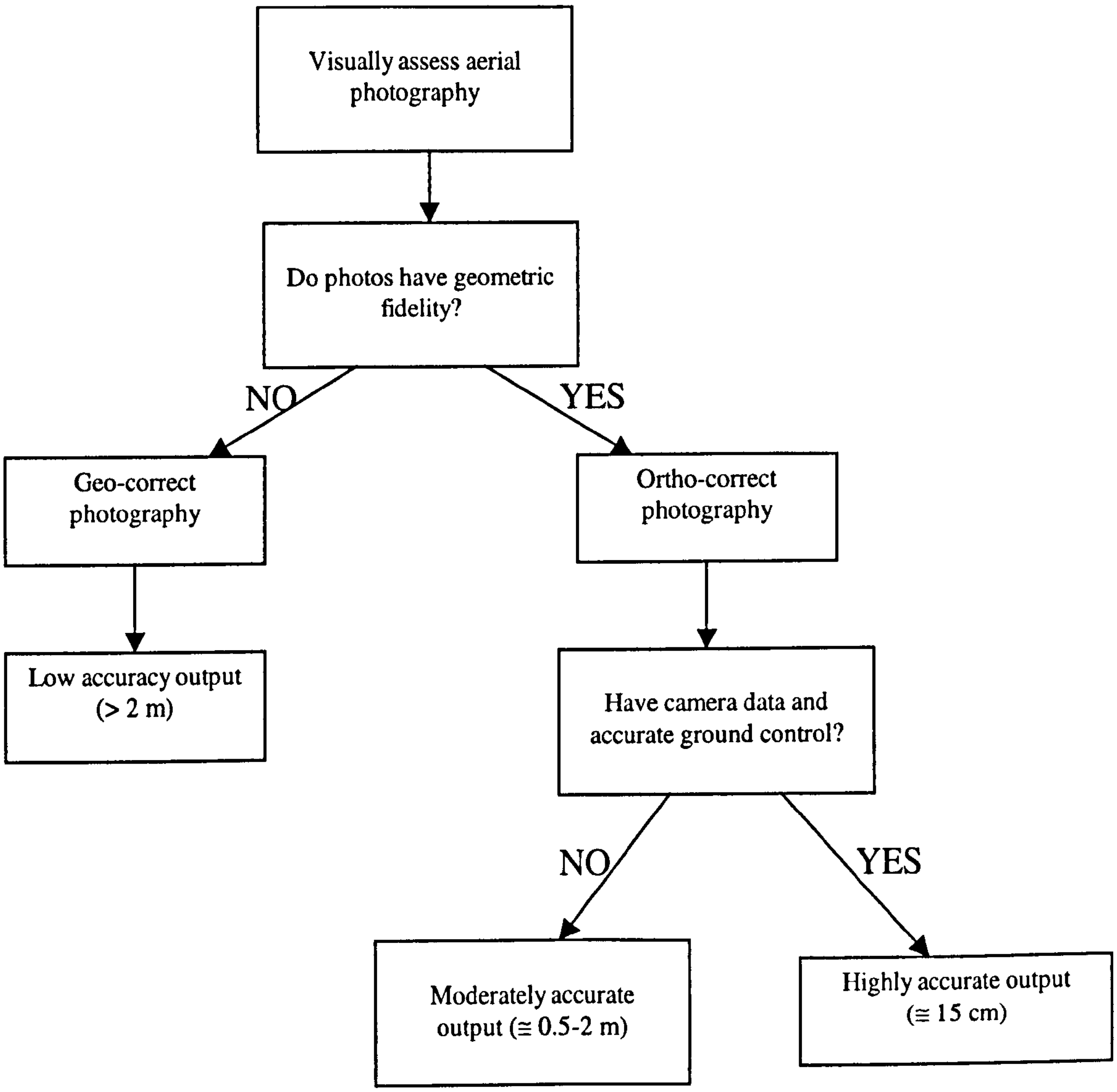
Bankfull position along the section line was identified on screen. Where location on screen was tenuous the original prints were viewed under a magnifying stereoscope to aid identification. Bankfull was easily identified (estimated to be within ± 1 pixel) where banks were near vertical because the break of slope was highlighted by the juxtaposition of dark pixels which represented water and lighter pixels indicative of floodplain vegetation. However, where banks were sloping the break of slope representative of bankfull was less easy to identify and relied more heavily on the use of a magnifying stereoscope. In such cases, the error in bankfull identification (or measurement error) was estimated as ± 3 pixels.

The pixel coordinate (shown on screen in Paintshop Pro version 5.01) of the intersection of the section line and bankfull was recorded for both left and right banks. Bankfull width was then extracted from the pixel coordinates using Pythagorean principles and knowledge of the pixel resolution. Using similar triangles the coordinate halfway between the two bankfull positions could also be calculated and compared for the different image dates. Again, Pythagorean principles were applied in order to calculate the lateral movement of the channel center along the section line between image dates.

6.4 2D measurements from rectified imagery in ERDAS Imagine – procedures

Whilst an estimate of measurement error is possible using the Paintshop Pro procedure, the error resulting from image distortion is not quantified. Consequently, it is impossible to fully determine the extent to which channel change recorded using the Paintshop Pro procedure is a function of image distortions rather than real morphological change. To combat this problem, and provide a comparative data set, analysis of channel change in the most active half of the Trefeglwys reach, between 13 and 14 km, was undertaken using geometrically corrected photography from which the total relative error values could be obtained. In addition to the air photo dates analysed previously, air photos from 1963 were also available for analysis and are therefore corrected and included.

Figure 6.2 Flow diagram of procedures used in air photo correction. After Mount et al. (2000)



6.4.1 Geo-correction

Geo-correction was applied to the 1948, 1963 and 1988 imagery using a second order polynomial geo-correction algorithm. The scanned images were imported into Imagine as raster layers and ground control was applied on screen. Ground control was derived from the 1:25,000 Ordnance survey map and bias was given to the identification of channel proximal ground features for use as ground control points (GCPs). GCPs were commonly the corners of fences and hedgerows or corners of buildings and bridges. A minimum of 20 GCPs were applied to each raster layer and these were then complemented with a minimum of a further 10 tie points. The correction algorithm was determined and applied and the resulting corrected raster layer was re-sampled to a 0.5 m pixel resolution.

6.4.2 Ortho-correction

The 1976 and 1995 imagery were both available as stereo pairs with geometric fidelity and as such were suitable for ortho-correction in OrthoMax. Ground control was provided by the 1:25,000 scale map and, as before, a minimum of 20 channel proximal GCPs were used and complemented by further tie points. The images were rectified to a transverse mercator projection using the default camera certificate. Root mean squared error (RMSE), representative of the absolute accuracy of the final rectified images, was greatest in the Z direction for both image dates (<5 m) and < 3.5 m in X and Y directions. Relative error between the 1976 and 1995 ortho-images was probably considerably lower (< 2 m in the X,Y and Z directions). Images were resolved to a 0.5 m pixel resolution prior to importing them into ERDAS Imagine as raster layers.

6.4.3 On screen checking

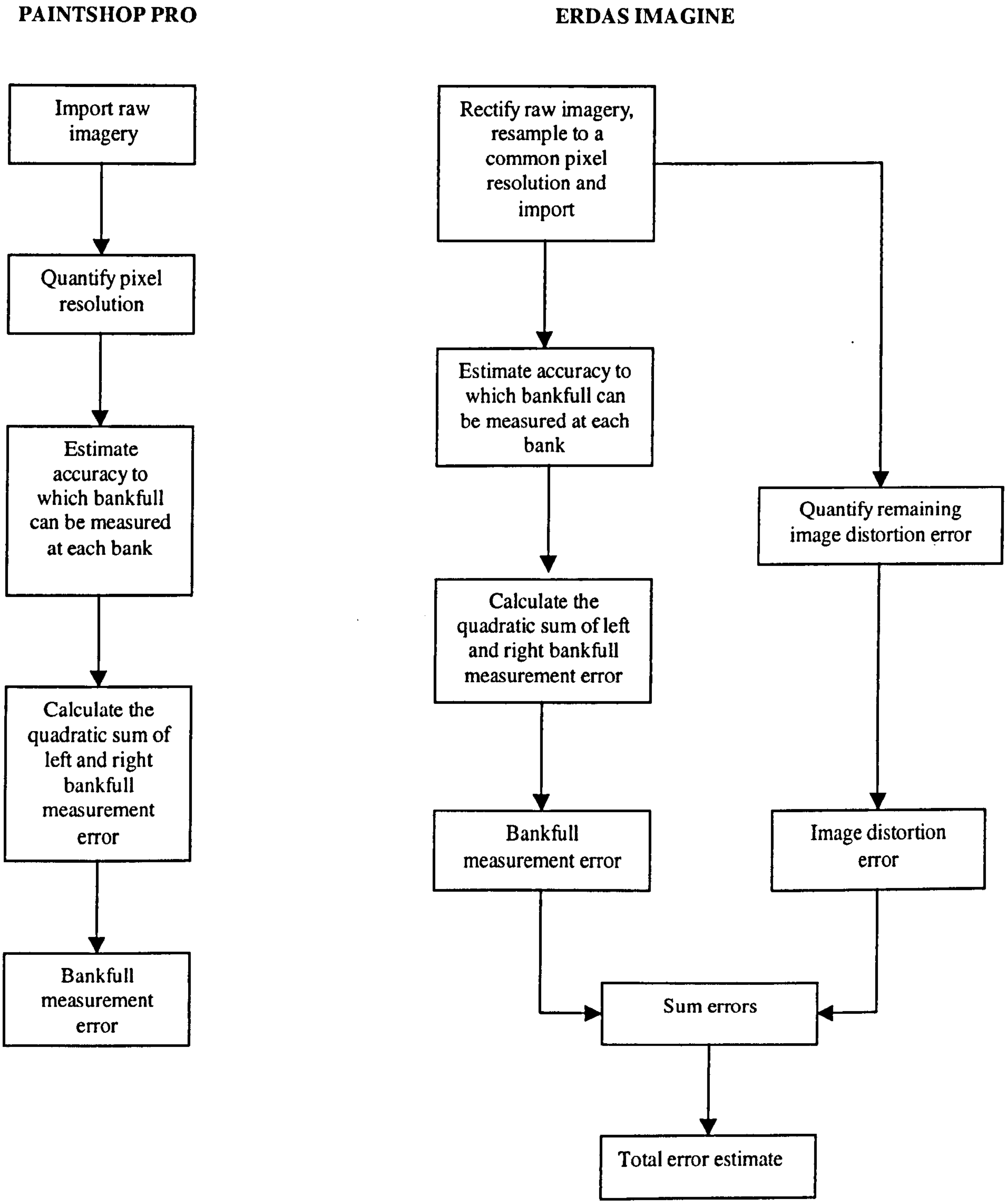
Rectified images were all corrected to GCPs, the coordinates of which were standardised to the Ordnance Survey National Grid. Consequently, the images could be geographically linked on screen in ERDAS Imagine. The swipe tool, which juxtaposes two images of different dates along a moveable swipe line according to the image coordinate system, was applied to the linked images. The accuracy of the match of channel proximal field boundaries across the swipe line was visually assessed. In this way a quick on screen check of the effectiveness of the correction procedures was achieved. A poor boundary match across the line would indicate a breakdown in the effectiveness of the correction algorithms. Channel proximal boundaries were seen to aligned well across the swipe; this was particular true for the 1976, 1988 and 1995 imagery, therefore the corrected images were accepted for analysis.

6.4.4 Measuring channel change

Concordant with the Paintshop Pro measurement procedures, a system of section lines was established on the rectified photography and the intersection of the section with left and right bankfull was recorded forming the basis of channel change analysis. To achieve this in ERDAS Imagine all image dates were geographically linked and section lines were drawn onto the 1995 image at 90 degrees to the channel using Imagine's measurement tool. In this way the line also appeared on all of the linked images. Section lines were established at a spacing of approximately 25 m. The precise downstream location of each section, relative to the upstream most section, was calculated by running the measurement tool along the middle of the channel. Where erosion or deposition between successive air photos had caused channel movement, relative to the 1995 air photo, section lines were adjusted to ensure that they were always orthogonal to the bank.

For each section and each of the linked images, the on screen pointer was positioned where bankfull and the section line intersected and the coordinates, displayed on screen, were recorded. Identification of bankfull on the 1976 and 1995 ortho-images was aided significantly by running the cross profile tool along the section line to obtain a channel cross-section on screen. For geo-corrected imagery use of the

Figure 6.3 Procedure for estimating error in historical aerial photograph analysis.



zoom function and the adjustment of image contrast and brightness aided bankfull identification. As a result less reliance on the original photography and the magnifying stereoscope was needed than with the Paintshop Pro procedure. Bankfull width was calculated from the bankfull coordinates using Pythagorean principles. Lateral change was calculated similarly, but using change in the position of the coordinate on the section line midway between left and right bankfull.

6.5 Measurement errors – comparing Paintshop Pro and ERDAS Imagine

The procedures used to quantify error in the Paintshop Pro and ERDAS Imagine techniques are summarised in figure 6.3. Image rectification is not possible in Paintshop Pro, therefore a simple estimate of bankfull measurement error is all that is possible. In ERDAS Imagine image rectification and the ability to geographically link images on screen means that in addition to a bankfull measurement error estimate, an image distortion estimate is also possible. The combination of the two errors provides a total error estimate for the ERDAS Imagine technique.

6.5.1 Paintshop Pro bankfull measurement error estimate

In evaluating measurement error it was assumed that bankfull, in all cases, was identified to an accuracy of ±3 pixels. This figure represents the maximum inaccuracy in bankfull identification on sloping banks, where a magnifying stereoscope was used to aid identification. The distance represented by 3 pixels in each of the image dates is presented in table 6.3.

The error inherent in the bankfull identification at each bank was non systematic and assumed to be subject to random uncertainties. Therefore the total error incurred in calculating bankfull width or channel center position for each sample date was given by the quadratic sum of the left and right bank error (Taylor, 1982):

$$\delta q = \sqrt{\left(\delta x^2\right) + \left(\delta y^2\right)}$$

(equation 6.1)

where

δq is the measurement error

δx is the error incurred in identifying left bankfull

δy is the error incurred in identifying right bankfull

Table 6.3 Measurement error at each bank and the quadratic sum of the error of both banks.

Image date	**Pixel resolution (m)**	**Accuracy to which left or right bank is identified (m)**	**Overall accuracy of bankfull width / channel center (m)**
1948	0.50	± 1.50	± 2.12
1963	0.50	± 1.50	± 2.12
1976	0.33	± 0.99	± 1.40
1988	0.35	± 1.05	± 1.48
1995	0.40	± 1.20	± 1.70

Rates of channel change are achieved by comparing quantifying differences in channel measurements from two image dates. The errors in quantifying bankfull location in each image are considered independent of each other and contain random uncertainties, therefore equation 6.1 was again applied with δx this time representing the overall measurement error from image x and δy representing the total measurement error from image y. The results are presented in table 6.4.

Table 6.4 Estimated measurement error between photographic dates.

Image dates	Measurement error (m)
1948-1963	±2.895
1948-1976	±2.535
1963-1976	±2.535
1976-1988	±2.028
1976-1995	±2.185
1988-1995	±2.247

6.5.2 Image and measurement error estimates for rectified photography in ERDAS

The Paintshop Pro procedure outlined previously allowed an estimate of bankfull measurement error to be calculated and applied to the data. However the absence of image rectification in the procedure meant that error resulting from distortion in the image could not be assessed. Here channel proximal image distortion error, after rectification, has also been estimated relative to the ortho-rectified 1995 image, and combined with an estimate of bankfull measurement error to provide an estimate of total error.

6.5.2.1 Estimating image distortion error

All of the rectified imagery was geographically linked on screen in ERDAS Imagine so that a line drawn onto one of the images would appear on all others. Near-channel features appearing on images of all dates (i.e. gateposts, the corners of fences and telegraph poles) were located and the distance measuring tool was used to calculate the distance from their location on the 1995 image to their location as they appeared on each of the other image dates in turn. In this way the positional error of the feature relative to the 1995 image was estimated for each image date in turn. 20 features spread throughout the study reach and located on both left and right banks were assessed and the maximum positional error recorded on each image date was assigned to represent the image distortion error. The maximum error was assigned because image distortions are commonly systematic and hence using the quadratic sum approach applied in section 6.2.3 is invalid. These errors are summarised in table 6.5.

Table 6.5 Image distortion error of rectified photography (i.e. the effectiveness of the rectification procedure).

Image date	**Rectification procedure**	**Location error relative to 1995 image (m)**
1948	Geo-rectified	± 1.00
1963	Geo-rectified	± 0.85
1976	Ortho-rectified	± 0.35
1988	Geo-rectified	± 0.55
1995	Ortho-rectified	± 0.00

6.5.2.2 Estimating bankfull measurement error

The extra functionality of ERDAS Imagine over Paintshop Pro (i.e. the ability to link images geographically, enlarge imagery quickly on screen and adjust raster layer contrast and brightness) meant that identification of bankfull was easier. It is estimated that, at worst, bankfull was identified to within ± 2 pixels (± 1 m). Moreover, where imagery was ortho-rectified the ability to call up and use cross profile plots in bankfull identification meant that bankfull could be located with improved accuracy. Therefore it is estimated that bankfull was identified to within ± 1 pixel (± 0.5 m) for the 1976 and 1995 imagery.

Overall error incurred in comparing channel change between two image dates (table 6.6) was obtained using the method outlined in section 6.3.2. This was then added to the quadratic sum of the image distortion error resulting from remaining post-rectification image distortion in the two image dates (see equation 6.1) to give an absolute error.

Table 6.6 Annual total error associated with photographic analysis within ERDAS Imagine using rectified imagery.

Study period	Quadratic sum of bankfull measurement error	Quadratic sum of image distortion error	Absolute error
1948-1963	± 2.82 m	± 1.31 m	± 4.13 m
1948-1976	± 2.23 m	± 1.06 m	± 3.29 m
1963-1976	± 2.23 m	± 0.91 m	± 3.14 m
1976-1995	± 2.23 m	± 0.35 m	± 2.58 m
1988-1995	± 2.23 m	± 0.55 m	± 2.78 m

6.6 Paintshop Pro and ERDAS Imagine: comparing error

The quantifiable measurement error in the Paintshop Pro and ERDAS Imagine procedures are presented in table 6.7.

Table 6.7 Summary of quantifiable error in Paintshop Pro and ERDAS Imagine aerial photograph analysis procedure

Image analysis period	Quadratic sum of bankfull measurement error in Paintshop Pro	Quadratic sum of bankfull measurement error in ERDAS Imagine	Total error (bankfull and image distortion) in ERDAS Imagine
1948-1963	± 2.895 m	± 2.82 m	± 4.13 m
1948-1976	± 2.535 m	± 2.23 m	± 3.29 m
1963-1976	± 2.535 m	± 2.23 m	± 3.14 m
1976-1995	± 2.185 m	± 2.23 m	± 2.58 m
1988-1995	± 2.247 m	± 2.23 m	± 2.78 m

Despite the increased functionality of ERDAS Imagine, the bankfull measurement error estimate becomes only slightly lower by using ERDAS Imagine to quantify bankfull position rather than Paintshop Pro. Indeed, the two methods appear roughly comparable in terms of bankfull measurement error estimates. The reason for this comparability lies in the pixel resolutions defined in the two techniques. Whilst the accuracy to which bankfull can be located in the rectified ERDAS imagery is improved (as low as ± 1 pixel in ERDAS compared to ± 3 pixels in Paintshop pro), the pixel resolution is coarser, being fixed by the resample size of 0.5 m rather than the image scale as in Paintshop Pro, which produces pixel resolutions as low as 0.33 m (1976). Consequently, the increased accuracy to which bankfull can be fixed in ERDAS is countered by the coarser pixel resolution.

Despite rectification of the ERDAS imagery, there is still some image distortion error inherent, however this is much lower in the higher quality, more recent photography. When this is added to the bankfull measurement error, the total error in the ERDAS imagine procedure becomes significantly greater than the bankfull measurement error in the Paintshop Pro procedure. However, the inability to rectify imagery within Paintshop Pro means that a comparison of the distortion error, and hence the total error estimate, between the two techniques can not be made. In order to overcome this, a direct measurement comparison, using identical section lines, was undertaken in the two software packages.

6.7 Raw and rectified imagery: comparing total error and combining data sets

To investigate the whether the accuracy of the Paintshop Pro and ERDAS imagine techniques were comparable, bankfull width at thirteen sections, identically located on the images, was calculated using both raw imagery in Paintshop Pro and rectified imagery in ERDAS Imagine. Analysis was undertaken around a meander bend where bankfull was easily identified, for the period 1948 to 1976 when error values for the rectified imagery are lowest. Using the meander bend ensured that the azimuth of the section lines on the photograph was highly variable. This ensured that the variable effect of direction on image distortion was taken into account (image distortion is not always constant in all directions). The 1948 imagery represents the poorest quality photography used in this study and the section of channel analysed appears on the margin of the print where lens and production distortion are maximum. Consequently, by analysing the 1948 imagery the impact of the highest magnitude distortions was examined.

The results of the analysis are shown in figure 6.4. The two data sets are seen to match well with a correlation coefficient of 0.95. The value of bankfull width calculated from raw imagery falls within the error bar range (± 3.36 m), associated with the use of rectified photography in ERDAS Imagine, for all sections except section 11. Here the difference is fairly small (1.48 m). At all sections the direction of channel response (either narrowing or widening) is the same. The results suggest that, in this study, the accuracy with which channel parameter values can be obtained using non-rectified imagery in Paintshop Pro is only slightly inferior to that possible using ERDAS Imagine and rectified imagery. This is largely due to the Trannon floodplain having low topographic relief – the major cause of image distortion, and the pixel resolution of the ERDAS Imagine imagery being coarser

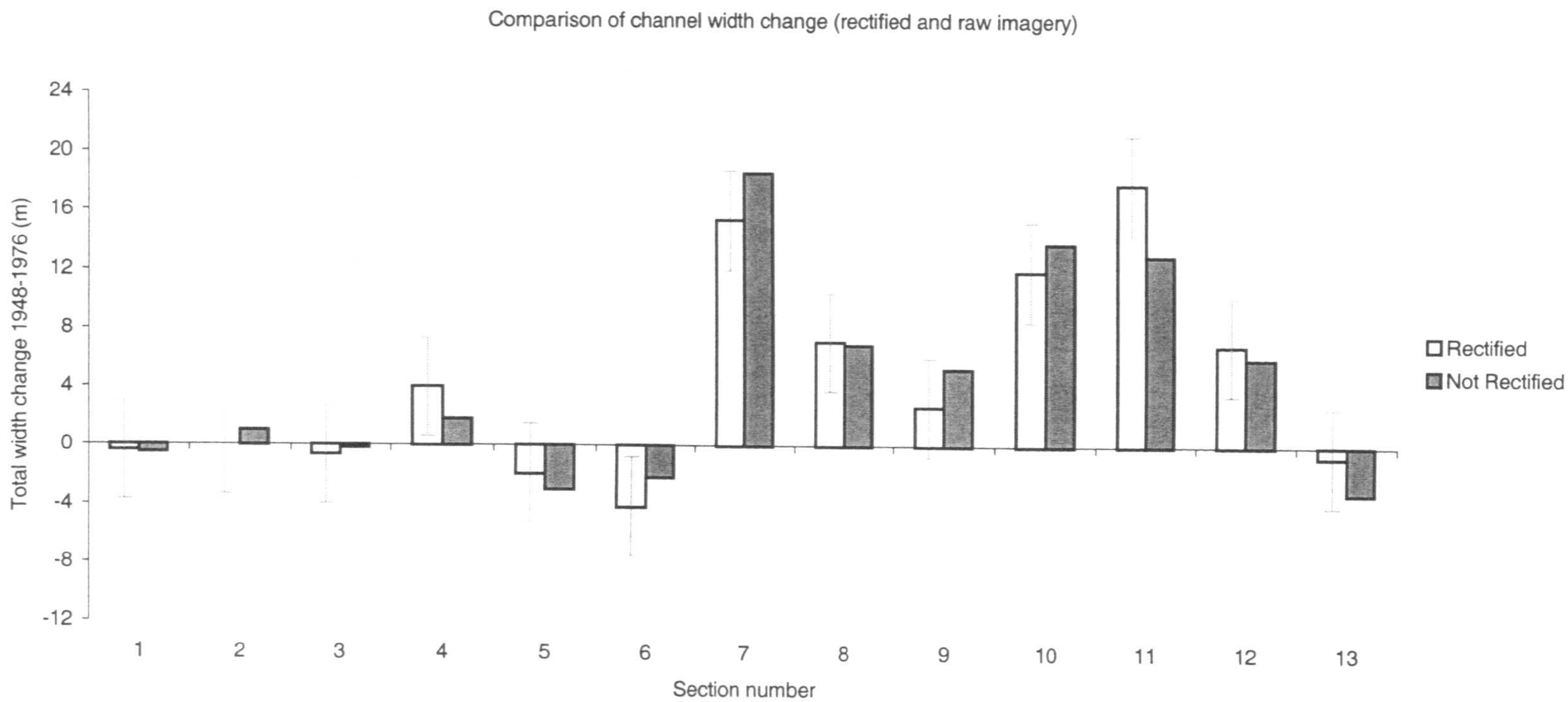


Figure 6.4 Comparison of bankfull width of 13 sections measured from raw and rectified imagery for the period 1948-1976.

than much of the Paintshop Pro imagery. The similarity does not imply that data generated using the Paintshop Pro procedures would be accurate in areas of substantial topographic relief (i.e. upstream of Llawy-y-Glyn) where image rectification would be essential prior to aerial photograph analysis.

Because of the comparable accuracy of the data from the two techniques the Trefeglwys data sets recorded using Paintshop Pro and ERDAS Imagine procedures have been combined. Also, the Llawr-y-Glyn reach data, obtained using the Paintshop Pro technique, is considered to have comparable accuracy due to the low topographic relief of the floodplain (channel gradient does not exceed 0.009) and it therefore also accepted for analysis. The total error estimates for analysis in ERDAS Imagine have been applied to all data in the analysis.

6.8 Llawr-y-Glyn study reach - results

Aerial photography covering the Llawr-y-Glyn reach was available for 1948, 1976 and 1995 only. As a result the rates of net change in channel width and lateral movement can only be investigated for the periods 1948-1976 and 1976-1995.

6.8.1 Llawr-y-Glyn reach – bankfull width change

In total 24 section lines were established throughout the 680 m long Llawr-y-Glyn reach, at a mean downstream spacing of 28.3 m. All sections were analysed for bankfull width change between 1948 and 1976, with 21 analysed between 1976 and 1995. 20 sections were analysed for lateral change in both periods, the shortfall in number being due to the obstruction of channel features by riparian vegetation. Summary statistics for bankfull width change at Llawr-y-Glyn are presented in table 6.8 and the data are presented graphically in figures 6.5 and 6.6.

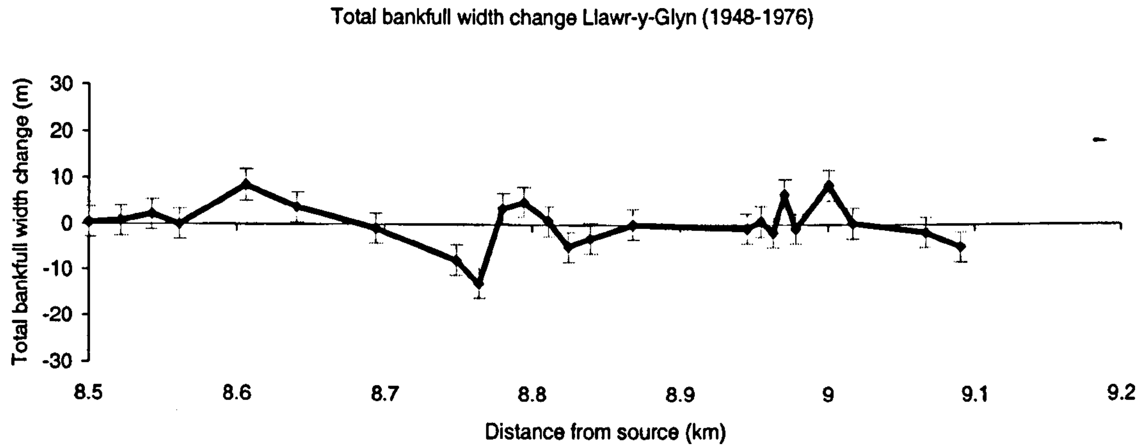
Table 6.8 Bankfull width change at Llawr-y-Glyn 1948-1976 and 1976-1995.

	Bankfull width change 1948-1976	**Bankfull width change 1976-1995**
Number of sections analyzed	24	21
Mean total width change between photos (m)	**0.008 ± 3.29**	**- 2.024 ± 2.58**
Maximum width increase between photos (m)	**8.4 ± 3.29**	**0.95 ± 2.58**
Maximum width decrease between photos (m)	**-12.9 ± 3.29**	**-7.8 ± 2.58**
Range between photos (m)	**21.3**	**8.93**
Mean width change ($m\ yr^{-1}$)	0.0003	-0.1065
Standard deviation ($m\ yr^{-1}$)	0.17	0.13

- 1948 to 1976:

11 sections displayed narrowing of the channel, 12 displayed an increase in width and 1 displayed no width change. However, only 10 sections (42 %) showed change greater than the error. Of these, 4 (17 %) showed narrowing and 6 (25 %) showed an increase in width. The maximum increase in width was 8.4 m ± 3.29 m, whereas the maximum decrease in width was of greater magnitude being –12.9 m ± 3.29 m. Despite these high change figures, the mean value is low due to the fact that both channel width increase and decrease occurred within the reach in similar proportion. The graph of total width change shows increases in width to be occurring in localised zones, spaced at approximately 200 m intervals. These can be seen at 8.6 km, 8.8 km and 9 km. Separating these zones are regions of no change or channel narrowing.

Converted to a mean time integrated change rate, the mean annual width change in the period is extremely low at less than 1 cm yr^{-1}. The maximum width increase and



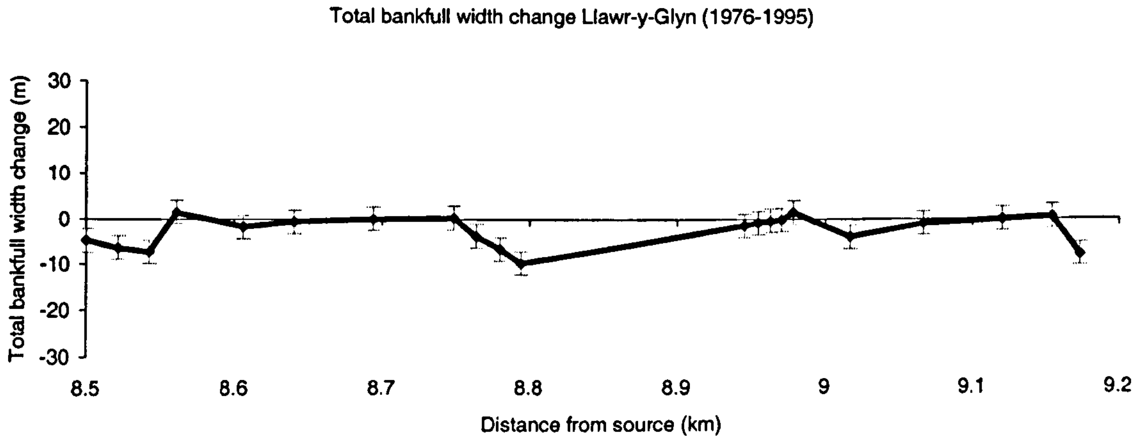
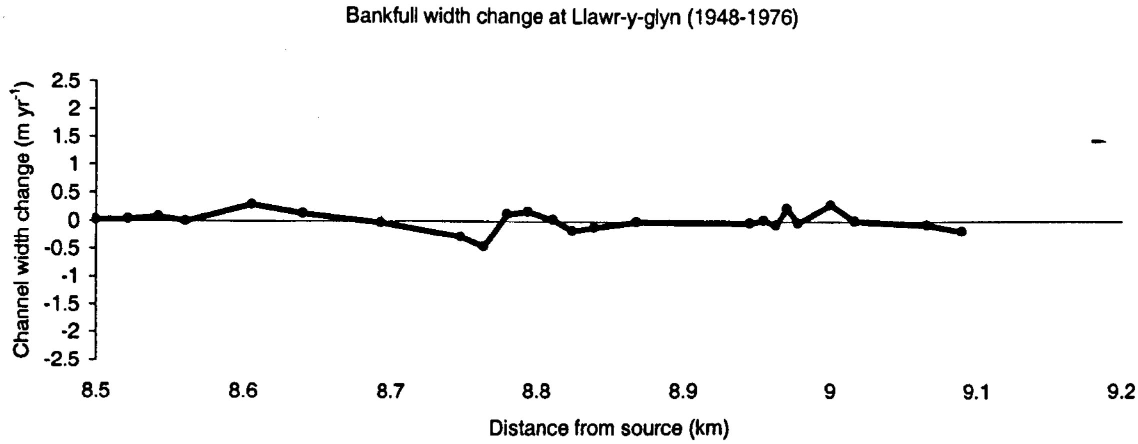


Figure 6.5 Total bankfull width change in the Llawr-y-Glyn study reach (1948-1976 and 1976-1995)



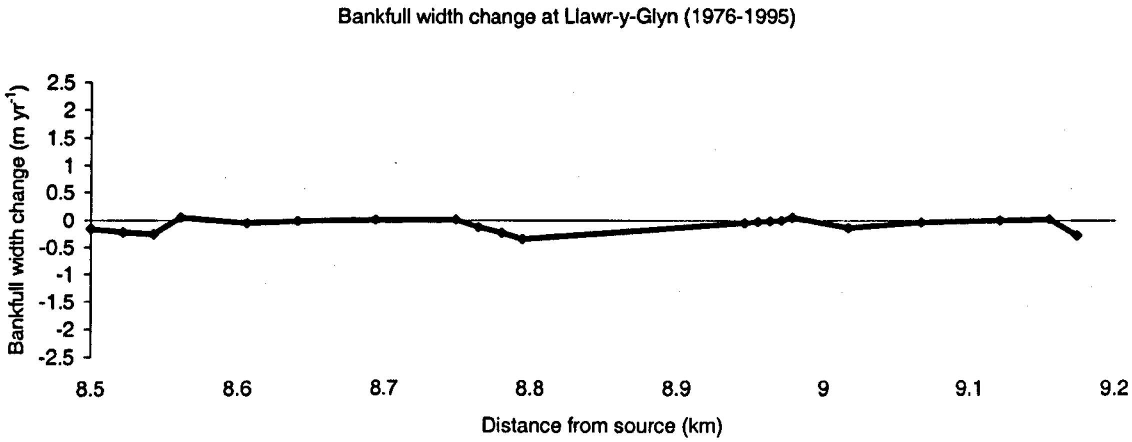


Figure 6.6 Mean time integrated bankfull width change in the Llawr-y-Glyn reach. (1948-1976 and 1976-1995)

decrease changes are much greater, being equal to 0.3 m yr^{-1} and -0.46 m yr^{-1} respectively, however these are still fairly low rates of change, as demonstrated when the data are displayed graphically (figure 6.6). The standard deviation reflects the variability in the response at each of the 24 sections being three orders of magnitude greater than the mean annual change rate.

- 1976 to 1995:

16 sections displayed a decrease in channel width, 4 showed an increase and 1 showed no change. Of these, 8 sections (38 %) showed change greater than the error, with all 8 displaying narrowing. This bias towards channel narrowing is reflected in the mean bankfull width change figure of –2.024 m ± 2.58 m; a value which nearly exceeds the error. The maximum decrease in width in the period is -7.8 m ± 2.58 m, slightly less than the maximum width decrease in the previous period. The total width change graph for the period shows the majority of channel narrowing to have occurred in a single zone lying between 8.75 km and 8.9 km.

Converted to a mean time integrated annual change rate, the reach sees a mean decrease in width of –0.1065 m yr^{-1} m – still a fairly low figure. The lower standard deviation of the data reflects the much smaller range of channel width response in this latter period, and this is visible when one compares the annual change rate graphs of the two periods.

6.8.2 Llawr-y-Glyn reach – lateral channel movement

Lateral channel movement summary data are presented in table 6.9. Total lateral movement and annual lateral movement graphs are presented in figures 6.7 and 6.8.

Table 6.9 Lateral channel movement at Llawr-y-Glyn 1948-1976 and 1976-1995.

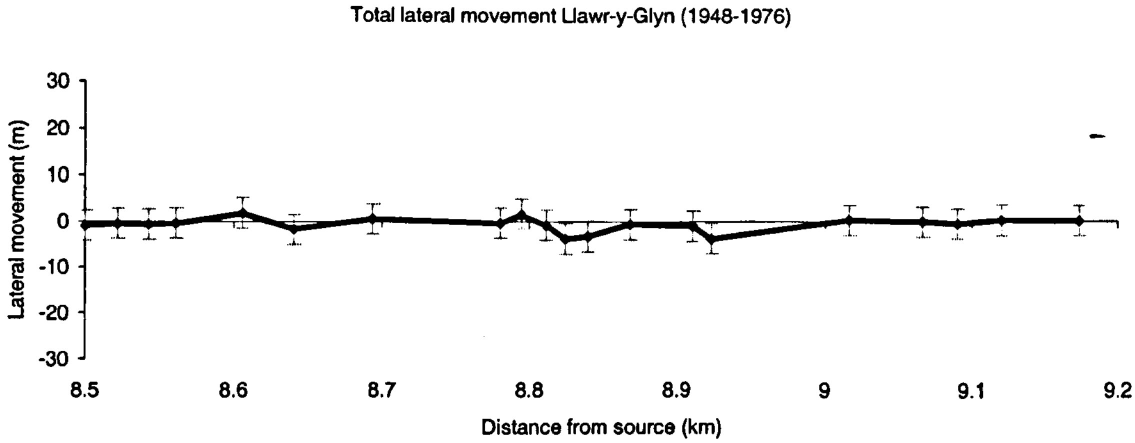
	Lateral channel movement 1948-1976	Lateral channel movement 1976-1995
Number of sections analyzed	20	20
Mean total lateral movement between photos (m)	**1.12 ± 3.29**	**1.33 ± 2.58**
Max. lateral movement between photos (m)	**3.92 ± 3.29**	**3.42 ± 2.58**
Min. lateral movement between photos (m)	**0.00 ± 3.29**	**0.00 ± 2.58**
Range between photos (m)	**3.92**	**3.42**
Mean lateral movement (m yr^{-1})	0.04	0.07
Standard deviation (m yr^{-1})	0.04	0.06

- 1948 to 1976:

Of the 20 sections analysed only 2 showed lateral movement exceeding the error (10 %). Maximum observed lateral movement was 3.92 m ± 3.29 m, equivalent to a mean time integrated annual rate of just 0.14 m yr^{-1}. The overall mean time integrated lateral movement rate is 0.04 m yr^{-1} and has a standard deviation of 0.04 m yr^{-1} which demonstrates that variability in the rate of channel movement between sections is extremely low. In summary, the Llawr-y-Glyn reach displays a very high level of lateral stability in this period.

- 1976-1995:

Of the 20 sections analysed 5 (25 %) showed lateral movement exceeding the error. The maximum observed movement being 3.42 m ± 2.58 m. This figure is 0.5 m lower than the maximum observed lateral movement in the previous period. In contrast, the mean total lateral movement is 1.33 m ± 2.58 m; 0.21 m greater than the previous period. This reflects the higher number of sections displaying lateral



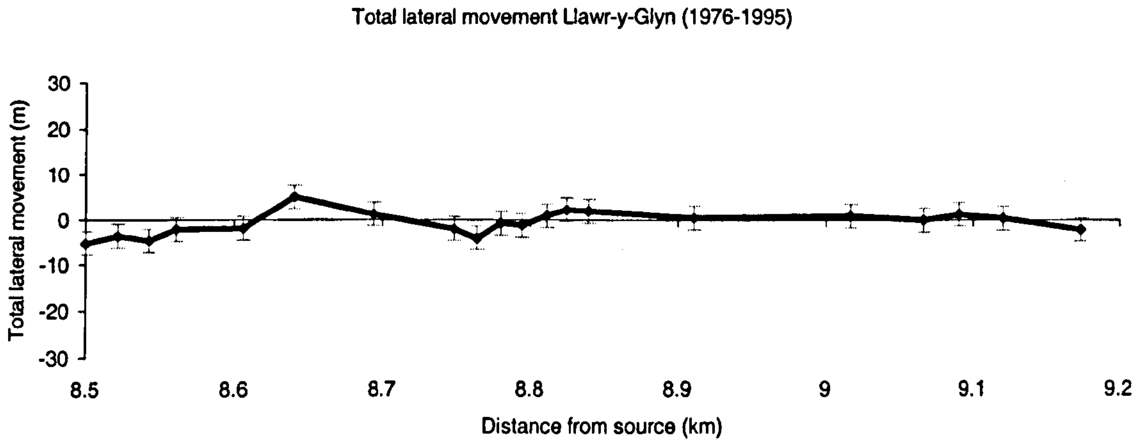


Figure 6.7 Total lateral movement in the Llawr-y-Glyn study reach
(1948-1976 and 1976-1995)
Positive values represent a left channel shift, whereas negative values represent a right channel shift.

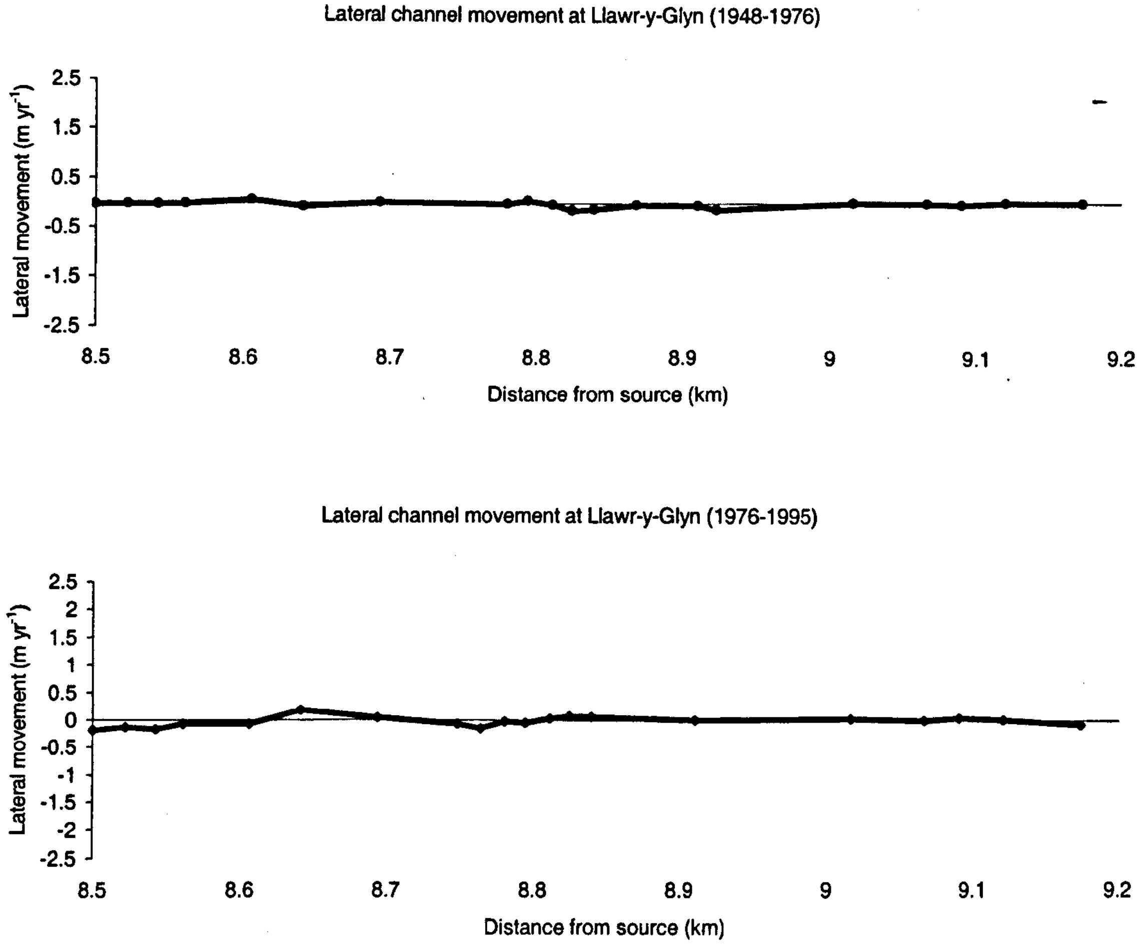


Figure 6.8 Mean time integrated lateral channel movement at Llawr-y-Glyn (1948-1976 and 1976-1995)
Positive values represent a left channel shift, whereas negative values represent a right channel shift.

movement. When the data are analysed in terms of an annual movement rate the reach is seen to have a level of lateral stability comparable to the 1948-1976 period. Mean rates of change are slightly higher (0.07 m yr^{-1}) as is the standard deviation, however the reach can be seen to display a laterally stable channel.

6.8.3 Llawr-y-Glyn –summary

The majority of historical change in the Afon Trannon at Llawr-y-Glyn can be seen to have occurred in adjustment of the channel width. A situation between 1948 and 1976 of localised zones displaying width increase and decrease, is replaced by a channel in which narrowing is dominant in the period 1976 to 1995. However, less than half of the sections display change greater than the error, and even the maximum width changes are relatively low when converted to a mean time integrated rate. The reach displays an extremely high level of lateral stability with very few of the section displaying lateral change exceeding the error and those that do exceeding it by < 1 m. In summary, the Llawr-y-Glyn reach has been largely stable between 1948 and 1995 with minor changes in bankfull width being the main feature of channel change.

6.9 Trefeglwys reach - results

For the periods 1948-1976 and 1988-1995, 131 section lines were established over the 2000 m Trefeglwys study reach at a mean spacing of 15 m. For the period 1963-1976 36 section lines were established in the upper 1000 m of the Trefeglwys study reach at a mean spacing of 27.8 m. No analysis of channel change was undertaken in the period 1976-1988 because the reach underwent extensive channel engineering in 1979, which consisted of altering channel width and depth by means of dredging the channel to a trapeziodal section. No major channel straightening was undertaken. Consequently, it is impossible to distinguish channel change, particularly in width, resulting from the channel engineering from change caused by variation in the sediment loads and streamflow characteristics of the river.

Between 1948 and 1976 all 131 sections were analysed for bankfull width change and lateral movement. However, increased riparian vegetation in the period 1988-1995 meant that only 118 and 116 sections were analysed for changes in width and lateral movement respectively. In the period 1963-1976 only ERDAS Imagine data are available for analysis, which for this period consisted of 36 sections in the upper 1000 m of the study reach.

6.9.1 Bankfull width change at Trefeglwys - results

Table 6.10 Bankfull width change in the Trefeglwys study reach

	1948-1976	1963-1976	1988-1995
Number of sections analysed	131	36	118
Mean total bankfull width change between photos (m)	**0.28 ± 3.29**	**2.47 ± 3.14**	**0.14 ± 2.78**
Maximum width increase between photos (m)	**17.4 ± 3.29**	**21.84 ± 3.14**	**13.72 ± 2.78**
Maximum width decrease between photos (m)	**-11.84 ± 3.29**	**-10.27 ± 3.14**	**-11.69 ± 2.78**
Range between photos (m)	**29.24**	**32.11**	**25.41**
Mean width change (m yr^{-1})	0.01	0.19	0.02
Standard deviation (m yr^{-1})	0.19	0.43	0.58

- 1948-1976:

In total, 67 % of the sections showed an increase in channel width and 31 % showed a decrease in width. The remainder showed no change. Of the 188 sections, 58 (44 %) of displayed width change greater than the total error with 35 showing channel widening and 23 showing channel narrowing. Mean total width change shows an overall increase in channel width and the range of width change observed is high at 29.24 m. Maximum increase in channel width for the period was 17.4 m ± 3.29 m

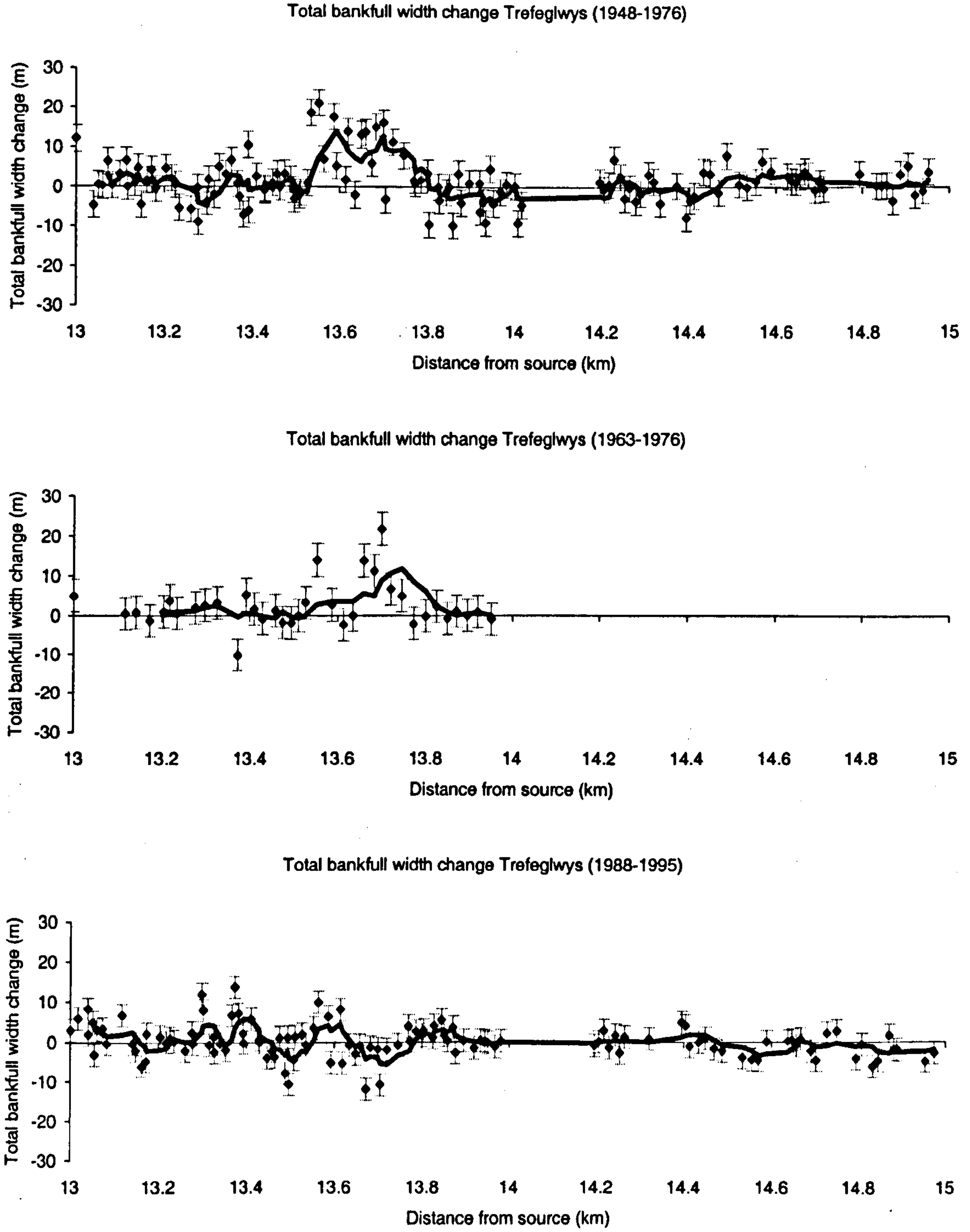
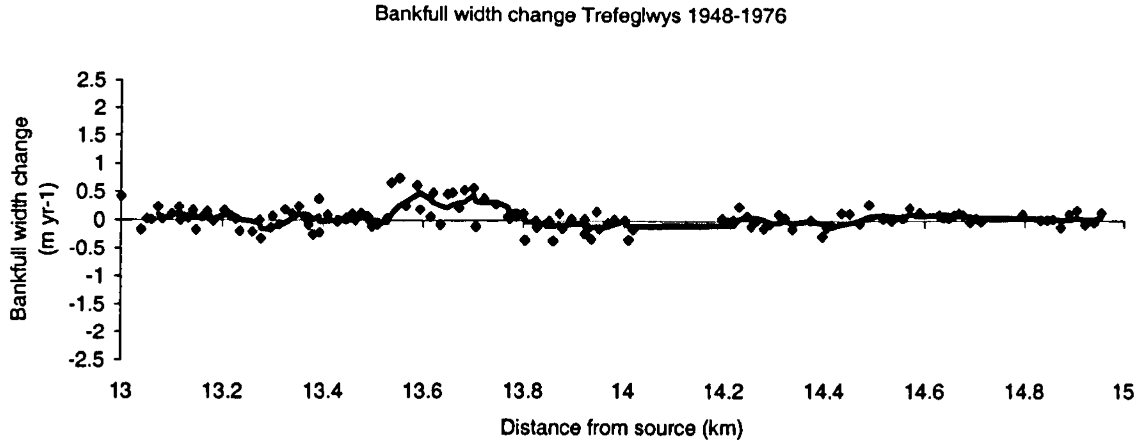
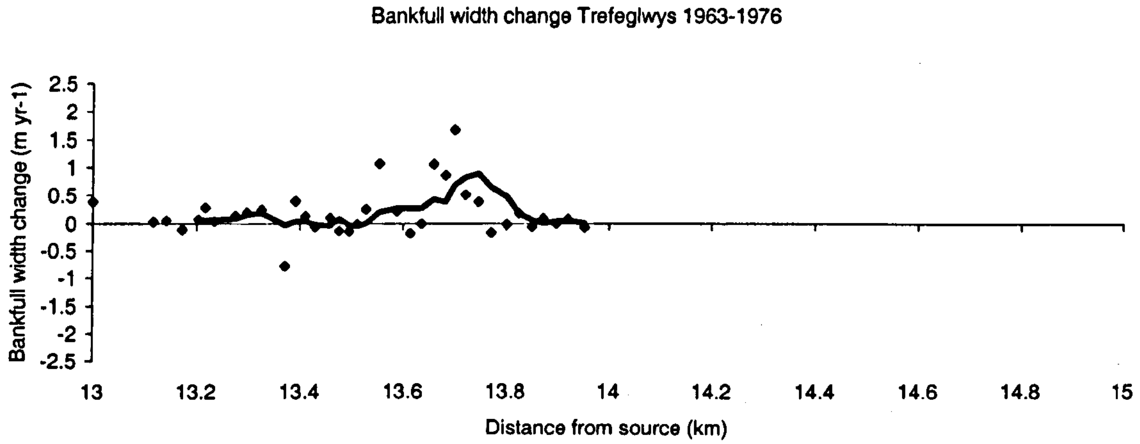


Figure 6.9 Total bankfull width change in the Trefeglwys study reach.
(1948-1976, 1963-1976 and 1988-1995)
A moving average trendline (n=5) has been fitted to remove noise
Note that no data exist between 14 and 14.2 km





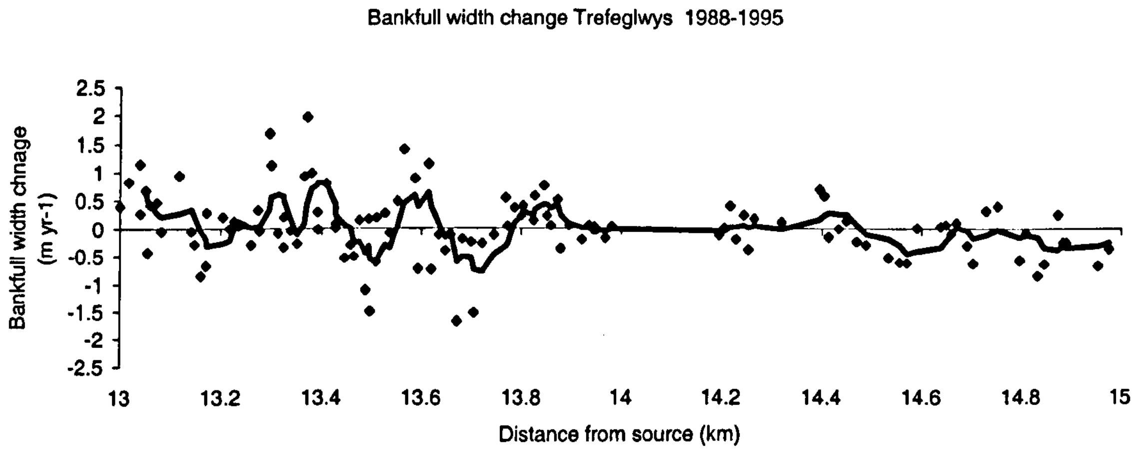


Figure 6.10 Mean time integrated bankfull width change in the Trefeglwys study reach. (1948-1976, 1963-1976 and 1988-1995)
A moving average trendline (n=5) has been fitted to remove noise

(equivalent to 0.62 m yr^{-1}) and exceeds the maximum narrowing which was less at 11.84 ± 3.29 m (equivalent to 0.42 m yr^{-1}).

Examination of the plots (figures 6.9 and 6.10) reveals a zone between 13.4 and 13.8 km in which the majority of bankfull widening occurred. This zone coincides with a meander bend about which extensive gravel deposition has occurred in the form of a point and lateral bar. Indeed, the upper 1000 m of the reach has generally experienced higher bankfull width change and a greater variability in channel response. Between 13 km and 14 km the channel sinuosity is $S = 2.1$, the range of total channel width change in the period is 30.76 m and the data have a standard deviation of 6.4 m. The mean width change is 1.87 m. Between 14 km and 15 km the channel sinuosity is lower at $S = 1.4$, the range of channel width change is almost half at 16.96 m and consequently the data have a lower standard deviation of 3.43 m. The mean width change is 0.55 m.

- 1963-1976:

The following results represent change in the upper 1000 m (13 – 14 km) only between 1963 and 1976.

Of the 36 sections analysed 23 showed channel widening, 12 showed channel narrowing and 1 showed no width change. Of these, 12 sections (33 %) showed width change greater than the total error with 11 showing widening and 1 showing narrowing. The mean width change in the reach was 2.47 m ± 3.14 m with a range of 32.11 m. Converted to a mean time integrated rate, the channel between 13 and 14 km experienced a mean increase in width of 0.19 m yr^{-1} and a standard deviation of 0.43 m yr^{-1}.

Compared to the bankfull width change between 13 and 14 km which occurred in the period 1948-1976, the channel can be seen to experience a higher mean width change in this period, although the mean change does not exceed the error in either periods. The range and standard deviation of width changes is slightly lower. This strongly suggests that the majority of the bankfull width change observed in the 28 year long 1948 to 1976 period actually occurred in the last 13 years, between 1963 and 1976. Indeed, this is reflected in the plots where the majority of bankfull width change can be seen to occur between 13.4 and 13.8 km in both periods, yet the mean time integrated rate is higher in the period 1963 to 1976.

- 1988-1995:

In total 118 sections were analysed throughout the 2000 m of the study reach. 61 sections showed widening, 56 showed narrowing and 1 showed no change. Of the 118 sections, 45 (38 %) showed change exceeding the total error and of these, 24 showed widening and 21 narrowing. The mean total width change showed a slight overall increase in width, however it does not exceed the error. The range of channel change is less than the period 1948 to 1976, however this is to be expected given the shorter time period between the photograph dates (7 years compared to 28 years).

As for 1948-1976, the plots (figures 6.9 and 6.10) show greater width change between 13 and 14 km than between 14 and 15 km. In the upper 1000 m of the reach the mean width increase was 0.67 m compared to a mean rate of channel narrowing of -0.87 m in the lower 1000 m. The range of width change observed between 13 and 14 km was more than twice that between 14 and 15 km, being 25.43 m and 10.73 m respectively. This higher variability in response is reflected in the standard deviation values. Standard deviation between 13 and 14 km = 4.50 m whereas between 14 and 15 km standard deviation = 2.62.

Comparing the results of channel width change between 13 and 14 km to the period 1963 to 1976 one sees an decline in the mean rate of width change from 2.47 m ± 3.14 m to 0.67 m ± 2.78 m. Converted to a mean time integrated rate, this represents nearly a four-fold decrease from 0.37 m yr^{-1} to 0.09 m yr^{-1}. However, neither figure exceeds the error and it is not possible to confirm that this response represents a real change. This decrease is not, however, accompanied by a decrease in the range of width changes observed (23.92 m for 1963 to 1976 and 25.43 m for 1988 to 1995) demonstrating that the downstream variability in channel width change has continued to increased over time between 13 and 14 km.

6.9.1.1 Summary

In all periods, mean bankfull width has increased with the maximum increase occurring between 13 and 14 km over the period 1963 to 1976. The majority of the width change seen between 1948 and 1976 is also apparent between 1963 and 1976, suggesting that little bankfull width change actually occurred between 1948 and 1963. However, mean bankfull width change does not exceed the error in any of the time periods and it does not, therefore, necessarily represent true change. What can be said though, is the percentage of sections experiencing change greater than the error has remained fairly consistent throughout the period (between 38 and 44 %). Moreover, the maximum increases and decreases in width have consistently exceeded the error. The maximum increases in channel width have consistently been of greater magnitude than the maximum decreases inferring a general increase in bankfull width. The more sinuous upper 1000 m of the study reach has experienced higher rates of bankfull width change than the lower 1000 m and a higher downstream variability in width changes.

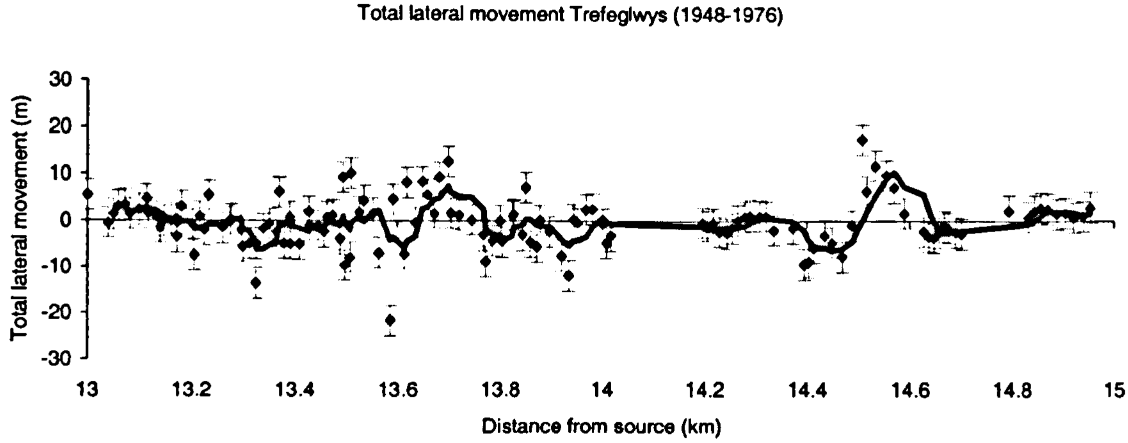
6.9.2 Lateral channel movement at Trefeglwys – results

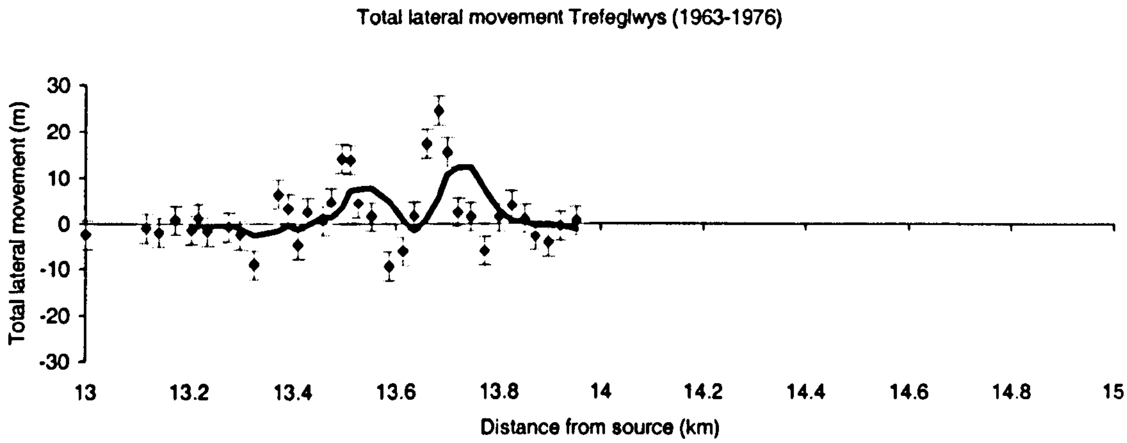
Table 6.11 Lateral channel movement in the Trefeglwys study reach.

	1948-1976	**1963-1976**	**1988-1995**
Number of sections analysed	131	36	116
Mean total lateral movement Between photos (m)	**3.64 ± 3.29**	**4.81 ± 3.14**	**2.73 ± 2.78**
Maximum lateral movement between photos (m)	**21.84 ± 3.29**	**24.44 ± 3.14**	**21.32 ± 2.78**
Minimum lateral movement between photos	**0.00 ± 3.29**	**0.52 ± 3.14**	**0.00 ± 2.78**
Range between photos (m)	**21.84**	**23.92**	**21.32**
Mean lateral movement ($m\ yr^{-1}$)	0.13	0.37	0.39
Standard deviation ($m\ yr^{-1}$)	0.13	0.43	0.35

- 1948-1976:

Of the 131 sections analysed, 51 (39 %) showed lateral movement in excess of the total error with a maximum lateral change of 21.84 m ± 3.29 m, equivalent to 0.78 m yr^{-1}. Despite only 39 % of sections exceeding the error, the mean rate of lateral change for the 2000 m reach is greater than the error at 3.64 m ± 3.29 m. As with bankfull width, the more sinuous upper 1000 m of the reach has experienced greater lateral change than the lower 1000 m. Between 13 and 14 km the mean lateral change in the period was 3.87 m ± 3.29 m and the maximum was 21.70 m ± 3.29 m, compared to a mean of 3.37 m ± 3.29 m and a maximum of 17.42 m ± 3.29 m between 14 and 15 km. Analysis of the plots (figures 6.11 and 6.12) shows the majority of lateral change to occur in two localised zones at 13.6-13.8 km and 14.5-14.7 km. These zones coincide with active meanders and associated point bars.





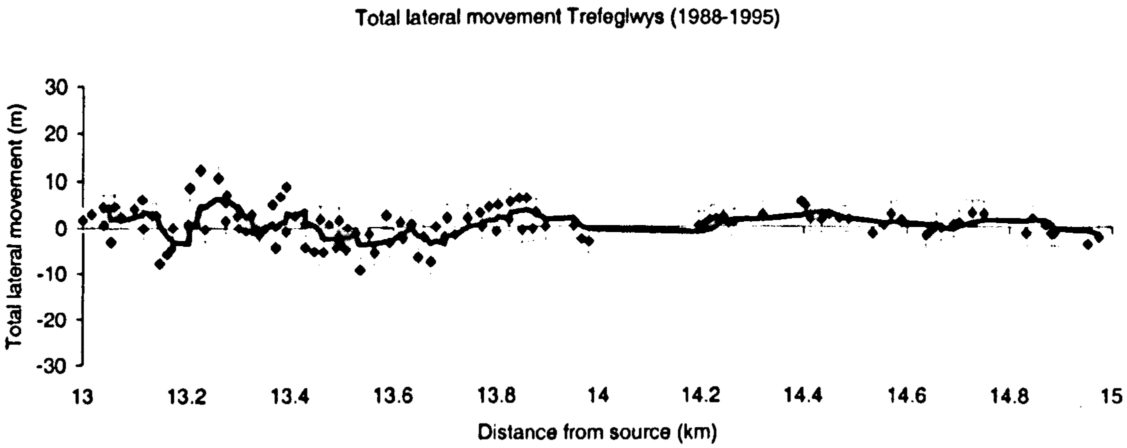


Figure 6.11 Total lateral channel movement in the Trefeglwys study reach. (1948-1976, 1963-1976 and 1988-1995)
A moving average trendline (n=5) has been fitted to remove noise
Negative values represent a left channel shift whereas positive values represent a right channel shift.

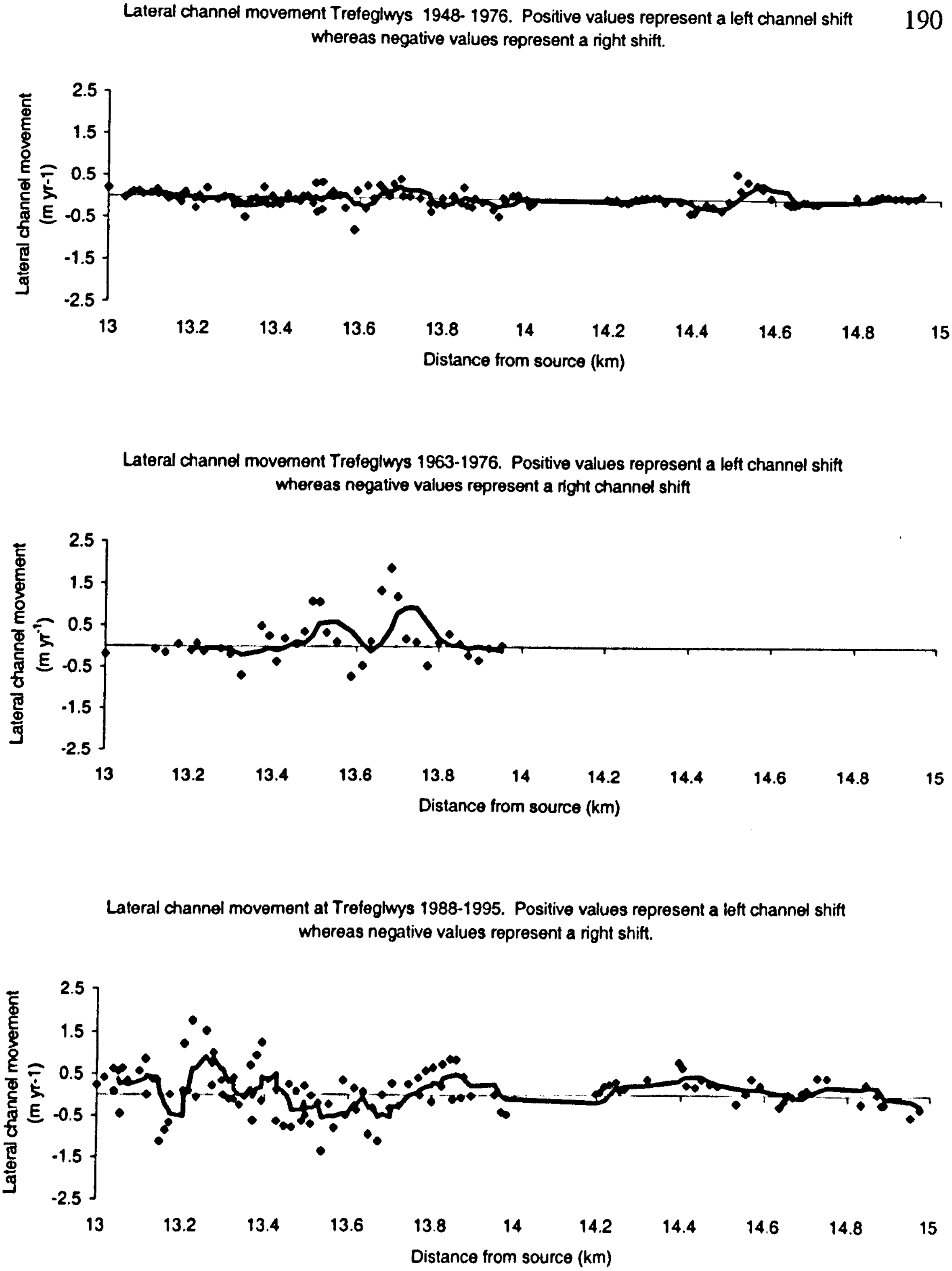


Figure 6.12 Mean time integrated lateral channel movement in the Trefeglwys study reach (1948-1976, 1963-1976 and 1988-1995)

A moving average trendline (n=5) has been fitted to remove noise

Negative values represent a left channel shift whereas positive values represent a right channel shift.

- 1963-1976:

The following results represent change in the upper 1000 m (13 – 14 km) only, between 1963 and 1976.

Of the 36 sections analysed, 16 (44 %) showed lateral movement greater than the error. However, mean lateral movement exceeded the error by more than 1.5 m. The maximum lateral movement was the highest recorded (24.44 m ± 3.14 m, equivalent to 1.88 m yr^{-1}).

Compared to the 1948-1976 results for 13 to 14 km, mean lateral movement is comparable for the two periods, with each lying within the error of the other. Comparison of the plots further demonstrates that lateral movement in the period 1963-1976 is similar to that in the period 1948-1976 which strongly suggests that the majority of lateral movement in the Trefeglwys reach occurred after 1963. This finding is in keeping with the bankfull width data for the same periods.

- 1988-1995:

Of the 116 sections analysed, 43 (37 %) showed lateral movement greater than the error. The mean lateral movement was slightly below the error (2.73 m ± 2.78 m), however the maximum lateral movement was greatly in excess of the total error (21.32 m ± 2.78 m, equivalent to 3.04 m yr^{-1}). Comparison of the plots reveals that total lateral movement in this period was the lowest observed, however when mean time integrated the mean lateral movement rate is similar to that of the period 1963 to 1976 between 13 and 14 km. The 1963-1976 mean rate of change between 13 and 14 km is 0.37 m yr^{-1} compared to 0.45 m yr^{-1} for the period 1988-1995. In keeping with the period 1948-1976, lateral movement between 14 and 15 km is lower than between 13 and 14 km (mean = 1.68 m ± 2.78, maximum = 5.22 m ± 2.78 between 14 and 15 km compared to mean = 3.19 m ± 2.78, maximum = 21.32 ± 2.78 between 13 and 14 km). Examination of the mean time integration plot between 13 and 13.8 km reveals a general pattern of reciprocating left and right channel shifts with a periodicity of approximately 200 m. This pattern coincides with the downstream location of migrating meander bends.

6.9.3 Summary

Mean total lateral movement in the Trefeglwys study reach has exceeded the error for the periods 1948-1976 and 1963-1976, however the data suggest that the majority of the lateral movement observed in the period 1948-1976 occurred after 1963. For all periods lateral channel movement is highest between 13 and 14 km and the maximum annual movement has increased from 1963-1976 (1.88 m yr^{-1}) to 1988-1995 (3.04 m yr^{-1}). In keeping with this increase in maximum lateral movement, mean lateral movement has also increased slightly in the upper 1000 m of the reach, from 0.37 m yr^{-1} between 1963-1976 to 0.45 m yr^{-1} between 1988-1995.

6.10 Llawr-y-Glyn and Trefeglwys reaches – channel change summary

Table 6.12 Summary of the main character and magnitudes of historical channel change in the Llawr-y-Glyn and Trefeglwys.

Reach / dates	Change magnitude (BFW = bankfull width, LM = lateral movement)		Change character
Llawr-y-Glyn 1948-1976	Mean BFW change (m)	0.008 ± 3.29	Extremely stable both in terms of BFW and LM.
	Mean BFW change (m yr^{-1})	0.0003	
	Max. BFW change (m yr^{-1})	0.3	
	Mean LM (m)	1.12 ± 3.29	
	Mean LM (m yr^{-1})	0.04	
	Max. LM (m yr^{-1})	0.14	
Trefeglwys 1948-1976	Mean BFW change (m)	0.28 ± 3.29	Low mean change in BFW and LM, but highly variable with high maximum changes. Majority of change occurred after 1963 and in upper 1000 m of reach.
	Mean BFW change (m yr^{-1})	0.01	
	Max. BFW change (m yr^{-1})	0.62	
	Mean LM (m)	3.64 ± 3.29	
	Mean LM (m yr^{-1})	0.13	
	Max. LM (m yr^{-1})	0.78	
Llawr-y-Glyn 1976-1995	Mean BFW change (m)	-2.02 ± 2.58	Laterally extremely stable. Some moderate reduction of BFW.
	Mean BFW change (m yr^{-1})	-0.10	
	Max. BFW change (m yr^{-1})	-0.41	
	Mean LM (m)	1.33 ± 2.58	
	Mean LM (m yr^{-1})	0.07	
	Max. LM (m yr^{-1})	0.18	
Trefeglwys 1988-1995	Mean BFW change (m)	0.14 ± 2.78	Laterally unstable with maximum LM an order of magnitude greater than observed previously. Variable BFW change with high maximum rates. Majority of change in upper 1000 m of reach.
	Mean BFW change (m yr^{-1})	0.02	
	Max. BFW change (m yr^{-1})	1.96	
	Mean LM (m)	2.73 ± 2.78	
	Mean LM (m yr^{-1})	0.39	
	Max. LM (m yr^{-1})	3.04	

From the above table it can be seen that the Llawr-y-Glyn and Trefeglwys study reaches have experienced very different levels of stability since 1948. The Llawr-y-Glyn reach has remained laterally stable throughout the entire study period, with low downstream variability. Bankfull width, whilst stable between 1948 and 1976, has increased subsequently with some sections experiencing moderate decreases in bankfull width between 1976 and 1995. However, the mean rate on bankfull width change has remained low.

At Trefeglwys the channel does not display the historic stability seen at Llawr-y-Glyn. This can be seen by comparing the mean time integrated bankfull width and lateral change plots from the two reaches (figures 6.5 – 6.12). Both maximum and mean rates of lateral channel movement have been consistently higher than at Llawr-y-Glyn with the maximum lateral movement rate in the period 1988-1995 being more than 16 times greater than the maximum lateral movement rate at Llawr-y-Glyn between 1976 and 1995. Maximum rates of bankfull width change have also been greater with consistent channel width increase observed at Trefeglwys compared to a decrease in channel width at Llawr-y-Glyn between 1976 and 1995. The downstream variability in bankfull width and lateral movement at Trefeglwys exceeds that of Llawr-y-Glyn and appears linked to channel planform, particularly meander bends.

6.11 Channel change prior to 1948

Leeks et al. (1988) examine the longer-term stability of the lowland Trannon, including the Trefeglwys study reach, for the period 1885-1976 using both historical maps and aerial photography. The methods by which this was achieved are not fully described, but it seems likely that the lack of a GIS in their methods would make the error inherent in their data at best comparable to that obtained using the Paintshop Pro technique previously described. Leeks et al. (1988) state that the lowland Trannon historically displays a 'moderate erosional environment' and demonstrate that bank erosion rates have increased significantly from 1901-1948 to

1948-1976, citing post 1948 upland afforestation as being of 'shorter-term relevance' to channel instability (Leeks et al. 1988, pg. 222). Compared to the data presented previously in this chapter, the findings of Leeks et al. (1988) strongly suggest that since 1948 the lowland Trannon has undergone extensive channel destabilisation. They calculate a maximum bank retreat between 1885 and 1948, prior to the establishment of Dolgau forest, of 0.3 m yr^{-1}. This is slightly less than the 0.37 m yr^{-1} observed between 1963 and 1976.

Examination of the 1963 photography, seemingly not analysed by Leeks et al. (1988), provides evidence of longer-term extensive channel destabilisation which challenges the purported view of a historically moderate erosional environment. This imagery was taken in the late afternoon, with the sun low in the sky, and snow cover is present. The combination of the low angle of the sun and snow cover means that shadows have been cast by very low topographic anomalies and these shadows are clearly visible contrasting greatly with the background snow. Examination of the aerial photograph shows a number of palaeochannels proximal to the main channel. The location of these palaeochannels and the current channel has been traced from the photograph and are presented in figure 6.13. Channel migration associated with meander development (marked A) and numerous small channel features (marked B), possibly associated with the confluence of the small tributary, are highlighted. The existence of these palaeochannels, of unknown age, infers that the Trannon has at some time been subject to substantial historical channel migration in the Trefeglwys reaches – evidence that suggests more than a 'moderate' historical erosional setting. However, it is not possible to confirm whether the migration has been a response to natural factors rather than man-induced factors. Interestingly further palaeochannels do not exist upstream of the Trefeglwys reach towards Llawr-y-Glyn, where stream gradient is higher, composite banks are not in evidence and the valley width is restricted. They are also not in evidence downstream of the Trefeglwys reach, where composite banks are replaced by sand / silt banks, until approximately 2 km upstream of the confluence of the Trannon and the Severn, a region known to be historically very active.

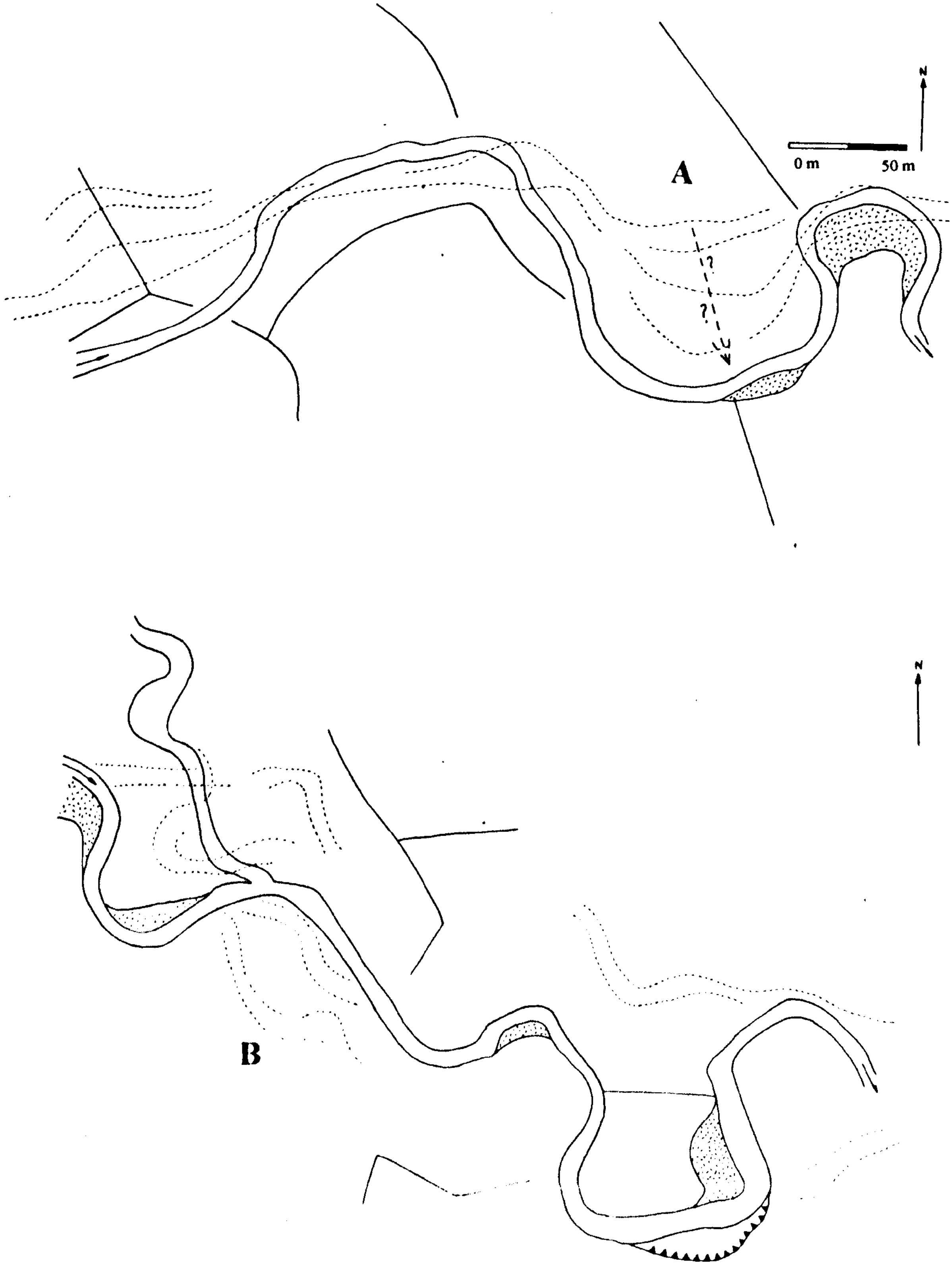


Figure 6.13 Location of identified palaeochannels (dashed lines) in the upper section of the Trefeglwys study reach from 1963 air photo. Single solid lines show field boundaries, hatched areas represent gravel bars and 'saw toothed' line shows location of inactive bank. The channel continues across the break in the diagram.

6.12 Medium-term response of gravel bar forms at Trefeglwys

6.12.1 Conceptual issues

In an attempt to identify the extent to which medium-term patterns of sediment deposition have influenced the medium-term patterns of channel change in the Trefeglwys study reach, gravel bar form areas have been quantified in the imagery of each date. The notion employed here is that an increase in sub-aerial gravel bar area implies an increase in the volume of gravel stored within the reach. However this is a simplistic approach which ignores the impact of fluctuating river stage and inconstant channel geometry on sub-aerial bar area. Inconstant river stage and channel geometry have been a particular problem in studies of braided streams (e.g. Carson and Griffiths, 1989; Sapozhnikov and Foufoula-Georgiou, 1996). The impacts are demonstrated in figure 6.14. Where discharge (Q) varies between image dates and the section geometry remains constant the resulting change in stage means that exposed gravel area (x) varies proportionally. In the case of figure 6.16a a drop in stage is accompanied by an increase in exposed bar without deposition. Similarly, where section geometry is not constant stage will vary, even in a condition of constant discharge. Also presented in figure 6.14 is the impact of scour which is accompanied by a drop in stage and an increase in exposed bar, even though discharge remains constant and there is no gravel deposition – indeed the section is degrading.

The above issues mean that in order to accurately infer volumes of stored gravel from exposed barforms quantified from aerial photography one would need detailed knowledge of the time that the image was collected, a discharge record covering that period and information of channel geometry. In this study the first two requirements can be met for the majority of the imagery, however the medium-term patterns of channel geometry are unknown. Therefore, the results presented here should be

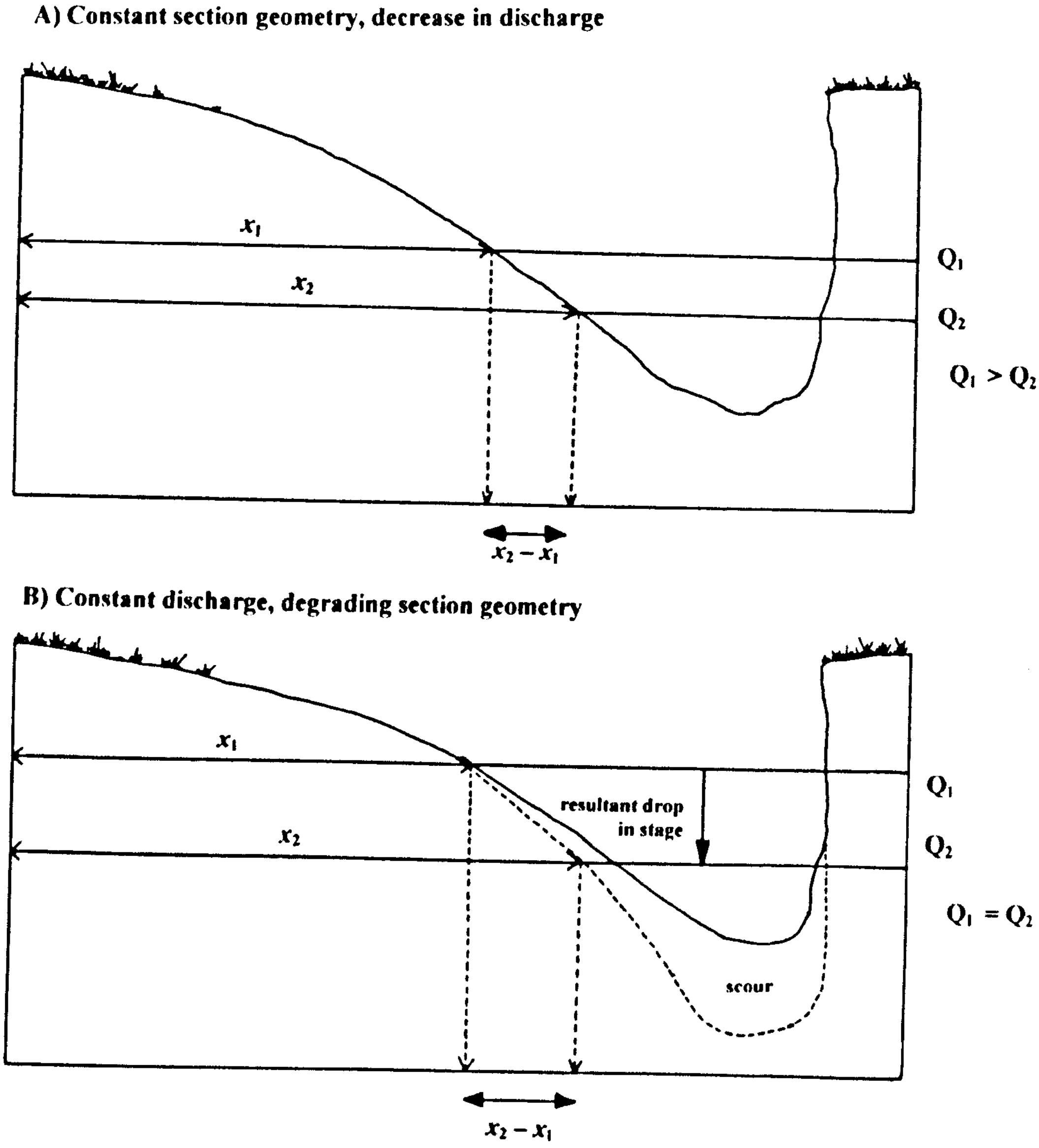


Figure 6.14 Schematic diagram demonstrating the impact of variable stage (Q) and section geometry on exposed gravel bar area (x). The numbers represent time with Q_1 indicating discharge at time t_1, and Q_2 indicating discharge at time t_2.

thought of only as indicators of stored gravel and may have extremely large associated error.

6.12.2 Procedure

Prepared and rectified imagery were imported into ERDAS Imagine as raster layers. The brightness and contrast of the images was adjusted to ensure that exposed gravel was easily discernable from surrounding vegetation and water. The edges of the gravel bars were digitised using the area measuring tool in ERDAS Imagine which quantified their sub-aerial surface areas. Where the edges of gravel bars were not obvious on screen, the original photographs were viewed under a magnifying stereoscope to confirm edge positions. Imagery from 1963 was not analysed because snow cover made identification of bar edges extremely difficult.

6.12.3 Results

Total exposed gravel bar area between 13 and 14 km downstream, for 1948, 1976, 1988 and 1995 are presented in figure 6.15. The exposed gravel bar area varies little apart from 1976 when the area is 1.4 times greater than that of 1948, the next largest. Examination of the modelled hydrologic record (see figure 5.2) reveals that 1976, a drought year, experienced particularly low flows. Daily mean stage on the 28.04.76 was 0.07 m, although this is only marginally lower than the daily mean stage for the 1988 and 1995 photography; 0.08 m and 0.10 m respectively. Hence, in terms of stage, the 1976, 1988 and 1995 data are comparable. No stage record exists for the 1948 image so the impact of stage on the data, and the extent to which direct comparisons can be made, are unknown.

Examination of the bankfull width change and lateral channel movement data, presented above, reveal that the peak in exposed gravel bar area coincides with the period of maximum medium-term channel change. The study reach experienced maximum morphological change in the period 1963-1976 compared to a more stable

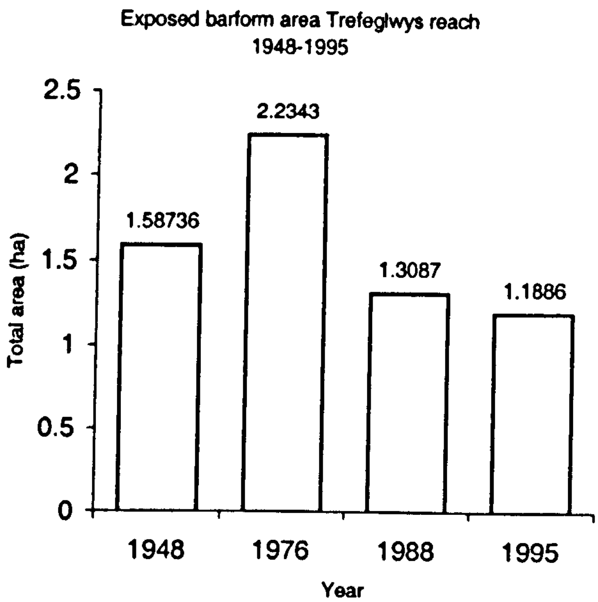


Figure 6.15 Total exposed gravel bar area (1948-1995)

channel prior to 1948 and in the period 1988-1995. This temporal association poses somewhat of a circular question: is the increase in gravel bar area a result of reduced stage due to channel section adjustment in the period 1963-1976, and not gravel deposition, or does the increase in gravel bar area represent genuine gravel deposition which has subsequently driven the 1963-1976 channel change? To answer this question fully would require the channel volume to be quantified for both dates. Attempts to derive high resolution DTMs for the channel from the available ortho-images (1976 and 1995), from which volumetric measurements can be obtained, have been unsuccessful because elevation changes within the channel were less than the RMSE associated with the DTMs. Therefore it is impossible to absolutely determine whether or not the peak in gravel bar area in 1976 represents an increase in gravel deposition.

7. CONTEMPORARY CHANNEL CHANGES: MAGNITUDE AND MECHANISMS OF CHANGE

7.1 Introduction

Within-channel sediment budgets form an important means of investigating the interaction which exists between bed load and channel change. Within a reach, bed load inputs (I) can be related to bed load outputs (O) by means of a simple equation:

$$I = \Delta S + O \qquad \text{(equation 7.1)}$$

where

ΔS is the change in sediment storage over time.

To produce a condition of equilibrium between bed load inputs and outputs, and assuming a condition whereby over bank deposition does not occur, one or a combination of the following conditions must be met:

1. All bed load entering the reach must be transported directly through the reach to the output without entering storage. Such a situation may occur if the study reach is very short. It may also occur if bed load transport rates are high and time between surveys is long. In such a scenario bed load may enter, move through and exit the reach in the inter-survey period.
2. The volume of bed load entering within-channel storage in the reach must be equaled by that scoured and transported out of the reach by within-channel erosion. This infers the overriding importance of bed load transport rates within the reach. In order to maintain input-output equilibrium, zones of low bed load transport must be compensated for by zones of high bed load transport. In this way substantial downstream change in bed morphology can occur whilst reach equilibrium is maintained.

3. The volume of bed load entering within channel storage in the reach must be equaled by the volume of material eroded from the banks. In this way the channel volume remains the same but the channel planform may undergo extensive change and may experience laterally instability.
4. The volume of scour and fill occurring in the inter-survey period balances.

Should none of these conditions be met, the volume of bed load within storage will change and this will be expressed by either aggradation or degradation identifiable as either an increase or decrease in the channel volume. Such fluctuations in stored bed load are common in gravel-bed rivers where bed load transport is normally capacity limited, and hence inextricably linked to the streamflow conditions. This chapter presents the methods and results used in the investigation of the within-channel sediment budget of the lowland Trannon. In particular the field studies aim to provide answers to the following questions:

1. Is significant bed load deposition and storage presently occurring in the Trefeglwys reaches? Is this situation different in the Ddranen Ddu reach?
2. What are the locations and magnitudes of bank erosion within the reaches? Do these patterns correspond with locations and magnitudes of bed load deposition?
3. What is the magnitude of bed load transport and deposition within the reaches? Can the passage of any bed load waves be identified in the study period?
4. What is the contribution to the bed load of the lowland channel from bank erosion? Is it more / less important a source than that from the upland catchment?

7.2 Channel surveying and topographic interpolation – background

The investigation of within-channel sediment dynamics necessarily requires temporally repetitive channel surveys capable of accurately representing the channel in three dimensions. However, the temporal and spatial resolution of these channel surveys requires careful selection and should be in tune with the temporal and spatial nature of change operating within the channel under study. In ephemeral streams two annual surveys, one before and one after flow occurs, may be acceptable to study large scale channel dynamics. However in perennial streams, with complex flood hydrographs, a higher temporal survey resolution may be considered desirable. Research concerned with extremely active channels (such as braided proglacial streams) commonly employ very high resolution procedures (24 hour temporal resolution and < 5 m spatial resolution) which operate over relatively short study periods, commonly weeks rather than years (e.g. Lane et al., 1995a). Conversely, in studies of extremely large scale channels lower resolution procedures are commonly adopted (e.g. Trimble 1983). This variability in approach represents the trade off between the necessary field time to repeatedly collect channel morphology data (as determined by the scale of the environment under study), and the minimum resolution required to successfully examine the parameters and processes of interest.

7.2.1 Repeat cross profiles – temporal and spatial resolution

Early studies of channel change were concerned mainly with planimetric adjustments of channels. Allen (1895), concerned with river bank erosion in Iowa, used planform surveys to monitor channel changes and planimetric re-survey methods have been used regularly since. Planform adjustments are a result of within-channel processes and more recently the research focus has shifted to understanding these processes. Consequently, planform surveys, where used, are normally undertaken in conjunction with other techniques capable of investigating the complex nature of within-channel processes.

Channel morphology is able to adjust both in the cross stream and in the downstream direction. Study of temporal changes in channel morphology therefore requires the ability to monitor the channel in the downstream (X), cross stream (Y) and channel depth (Z) directions over regular time intervals. Normally this has been achieved via a network of sub-parallel benchmarked cross profiles, orthogonal to the channel banks, spaced at fixed intervals downstream and re-surveyed periodically (e.g.

Lawler 1993; Wathen and Hoey 1997; Nicholas and Sambrook-Smith 1998). It is standard to have a fixed measurement interval in the cross stream direction. The volume of the channel can be easily determined via simple multiplication of the section area, by an amount equaling half the section spacing, in both an upstream and downstream direction. Analysis of the temporal change in channel volume combined with knowledge of sediment loads and streamflow conditions allows investigation of complex and spatially variable within-channel processes. The technique has been shown to work well over straight reaches, however where wandering and meandering channels are of interest it is not possible to maintain parallel sections which are orthogonal to the bank. In such situations neighboring section lines are non parallel resulting in enormously complex calculations, or a simple 'fudge' in order to calculate channel volume. This problem explains the over representation of straight reach studies in the literature. In this study the problem of non parallel cross profile lines is overcome by the processing and surface interpolation of field data within a GIS. This approach removes the manual calculation complexity associated with non parallel section lines.

In studies which use repeat cross profiles a necessary trade off exists between the time spent collecting data points and the frequency of return to the profiles which is logistically possible. Such a trade off is particularly apparent in studies of morphologically highly active channels. Consequently, important questions have been raised relating to survey resolutions. Lane et al. (1995a) point out that it is common for cross profile surveys to have a far higher density of points in the cross-stream direction than the downstream. Indeed, Ferguson (1992) uses a downstream spacing of 20 m, Goff and Ashmore (1994) use a spacing of 10 m and Ferguson and Ashworth (1992) use a spacing of 5m yet all of these studies use a far higher survey resolution (i.e. 0.5 m) in the cross-stream direction. Lane argues that whilst the cross-stream survey density is commonly acceptable, large distances between survey lines in the downstream direction are only acceptable in channels where morphologic change occurs mainly laterally. As a guideline Lane et al. (1994) suggest an appropriate downstream spacing of 2-3 m. However, their study is concerned with a small proglacial, braided stream which experiences high rates of morphologic change in both the cross and downstream directions. In larger, meandering rivers, where the rate of morphologic change is less, the cross profile spacing can probably be safely extended, although as yet no generally accepted rules pertaining to the error incurred in such extension exist.

The temporal resolution of re-survey reported in the literature varies with the fluvial environment of interest. In order to obtain the highest possible data resolution surveys should be repeated after every flow event above the threshold at which morphological change is achieved. In the case of proglacial streams, where peak discharge occurs diurnally, this is at least once every 24 hours – although many proglacial studies fail to achieve such resolution (cf. Ashworth and Ferguson, 1986; Powell and Ashworth, 1995). In the case of temporally irregular, compound flood peaks (such as those which occur in the Afon Trannon) it is impossible to predict the timing of high flows above the threshold of morphologic change. Consequently, the resolution of re-surveys is more arbitrary for such rivers.

7.2.2 Surface interpolation from surveyed cross-profiles

Recent developments in digital photogrammetry and surface interpolation software housed within Geographic Information Systems (GIS) have allowed advancements in quantifying and examining patterns of morphologic change, although the issues pertaining to survey resolution remain.

It is possible for surfaces to be interpolated between spatially auto-correlated points to form digital terrain models (DTMs). These allow estimates of topographic values to be generated where no measurements exist based upon proximal known values (Lam, 1983). The application of surface interpolation algorithms to repeat cross profile data sets solves several problems:

1. Channel volume can be calculated from non parallel section lines allowing investigations of meandering reaches; indeed surveys do not need to be constructed along predefined cross profile lines
2. The spacing of points along a section line need not be constant, nor in the same location for every re-survey allowing each survey to better match the section shape
3. The number of points along each cross profile need not be constant with successive surveys
4. Interpolated surfaces can be converted to contour based images allowing links between overall volume change and bedform evolution to be examined
5. Successive DTMs may be subtracted from one another to show temporal changes in the distribution and magnitude of bed load scour and deposition
6. Both planform and cross profile surveys may be included together in the interpolation procedure to enhance survey point density and to improve channel boundary definition

Surface interpolation algorithms require every survey point to possess spatial reference for X,Y and Z directions, meaning that field collection of data is best achieved using either high quality surveying equipment - e.g. a total station capable of electronic distance measurement (e.g. Lawler, 1993; Lane et al., 1998) or photogrammetric techniques (e.g. Lane et al., 1994). The major drawbacks of photogrametric determination of channel morphology are the current problems associated with being able to 'see through water' (Westaway et al. 2000) and the difficulty of obtaining air-photography or other data sources of sufficiently high resolution and that have been acquired frequently enough. Consequently DTMs can be simply constructed only for features above the water line – arguably the least interesting region of the channel in terms of morphologic change. As a result most studies concerned with reach and sub-reach scale channel morphology use data collected using a total station. Lane et al., (1994) attempted to overcome the limitation of photogrammetry in respect to seeing through water by integrating both terrestrial photogrammetry and manual survey data to reconstruct channel morphology. They report considerable success in this approach, however their study is of a small proglacial stream reach (50 m in length) where there is no riparian vegetation which obscures the channel from terrestrial photography. In the case of this study, which has assessed a combined reach length of approximately 530 m, and where the channel is commonly obscured by riparian vegetation, the use of terrestrial photogrammetry is seriously limited. Consequently, the channel models generated and analysed here are produced entirely from repeated channel cross-profiles.

7.3 Channel surveying and topographic interpolation – procedures

7.3.1 Cross profile surveys

Surveys were achieved using a Leica TC600 total station in conjunction with a WILD GPR1 reflecting prism, although a Leica NA20 quick set level and standard surveying staff were used occasionally following equipment malfunction.

Repeat cross profile lines were established at each of the three study reaches and tied into a fixed survey of total station location benchmarks marked by wooden stakes. The survey was not tied into the Ordnance Survey National Grid. North for each re-survey was set with a Silva type 15 compass clinometer with an error of < +/- 1°.

Cross profile lines were established by fixing a taught length of cord between the start and end pegs. In total 69 section lines were established through 3 study reaches. The locations of the three study reaches are shown in figure 7.1 and the non-parallel cross profile setting at the Chapel View (CV) and Bodiach Hall (BH) sites are shown in figure 7.2. The cross profile setting in all study reaches is further summarised in table 7.1 below:

Table 7.1 Summary of cross profile spacing at the field study sites.

Reach	Number of Sections	Section Spacing
Chapel View (CV)	CV1-CV12 (12)	<10 m – 24 m. Non Parallel
Bodiach Hall (BH)	BH1-BH47 (47)	<4 m – 21 m. Non Parallel
Ddranen Ddu (DD)	DD1-DD10 (10)	<4 m – 6 m. Parallel

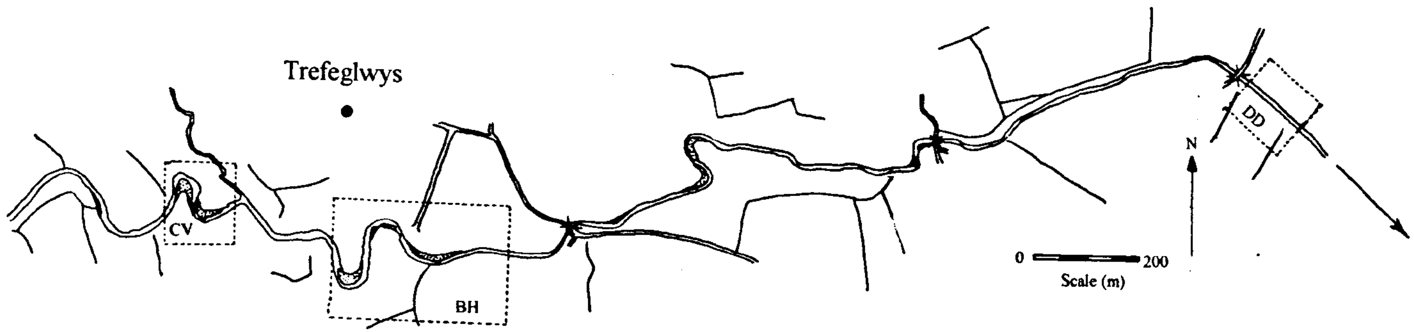


Figure 7.1 Location of the Chapel View (CV), Bodiach Hall (BH) and Ddranen Ddu (DD) reaches in the lowland Afon Trannon.

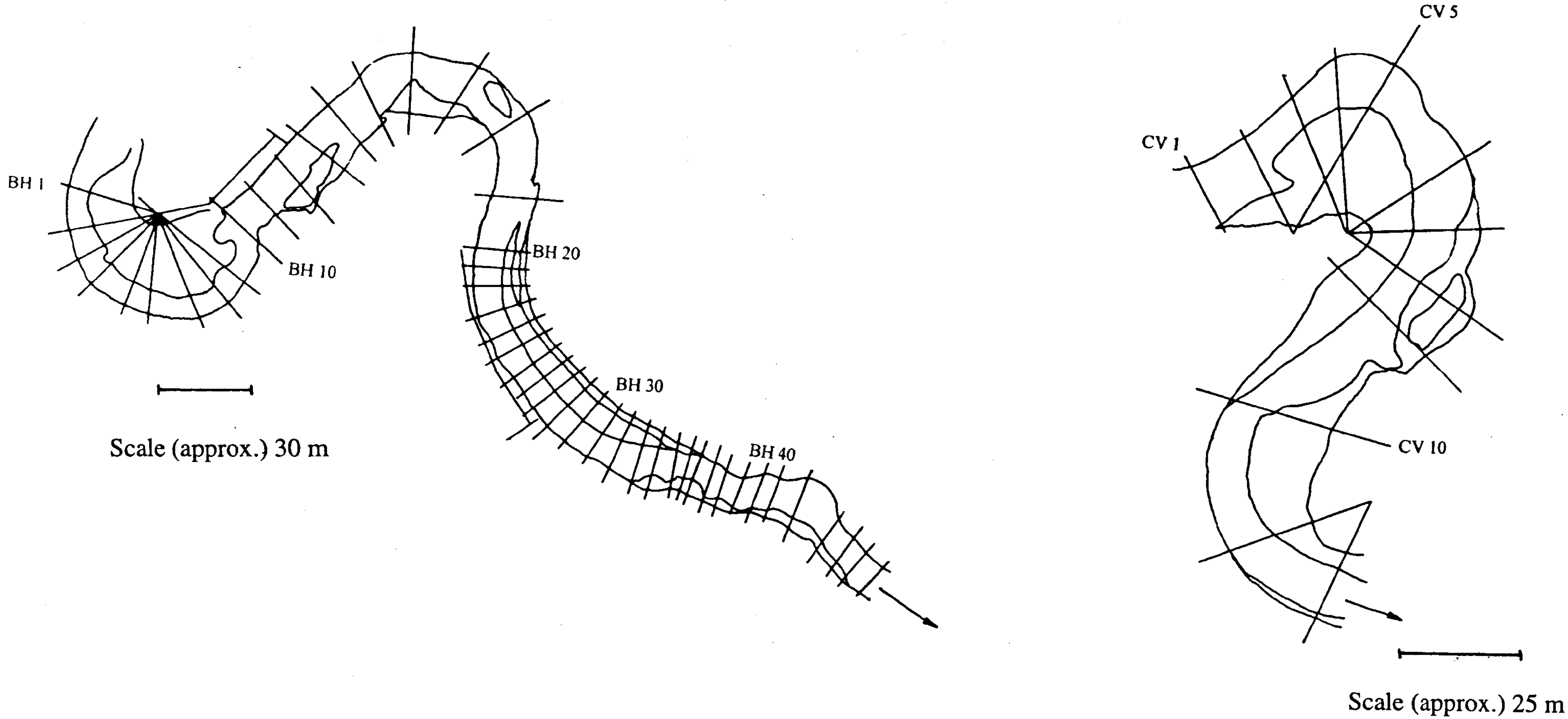


Figure 7.2 Position of cross profile lines at the Bodiach Hall (BH) and Chapel View (CV) reaches.

7.3.1.1 Section Spacing / Point Resolution

The cross profile spacing used was variable, ranging approximately between 4 m and 24 m. Where downstream channel morphology appeared complex, downstream resolution was increased. Obscured section lines due to vegetation forced some increases in downstream section spacing. The meandering nature of reaches CV and BH required a non-parallel cross profile setup to ensure profile lines were always orthogonal to channel banks. Where significant lateral channel change occurred at meander bends during the study, section lines occasionally ceased to be orthogonal to the bank (a problem noted by Lawler, 1993). In the air-photo analysis, described in chapter 6, this impact of lateral channel movement on the channel width and channel center location was corrected by section rotation. However, here the use of a GIS to plot and interpolate the section data to obtain channel volume information meant that section rotation was unnecessary because once interpolated, channel information could be gained from a section line drawn onto the interpolated surface at any location, unrestricted by the original cross profile azimuth.

Cross stream spacing was determined by the topography of the profile line and was therefore not constant. Where topographic change was irregular in the cross stream direction, point density was increased to better represent the surface. Conversely, where topographic change was regular (i.e. flat gravel bar surfaces and floodplains) point spacing was increased. The number of points surveyed across each profile varied from survey to survey depending on the topographic nature of the bed at the time of survey and ranged from between 25 and 50.

7.3.1.2 Pegging Procedure

Cross profile start and end points were marked with painted, numbered wooden stakes. The coordinates of each peg relative to the total station benchmarks were recorded allowing accurate relocation (to within 0.05 m in any direction) in case of loss. Start and end pegs were commonly located at least 5 m from bankfull, although the distance was less in the laterally stable Ddranen Ddu reach.

7.3.2 Planform Surveys

Planform surveys were conducted at every re-survey for the BH reach and all but re-survey 5 for the CV reach. Planform surveys were undertaken using the total station set at the same benchmarks used for the cross-profiles. Points were commonly < 5 m apart and followed continuous morphological features. Every planform records the following channel morphology parameters:

1. Bankfull (located in accordance with Osterkamp and Hedman (1982) as the point at which any increase in bank elevation is accompanied by an extremely large cross-channel distance)
2. Bank base (vertical and near vertical banks only)
3. Active gravel barforms (including unit bars)
4. Waterline
5. Active channel

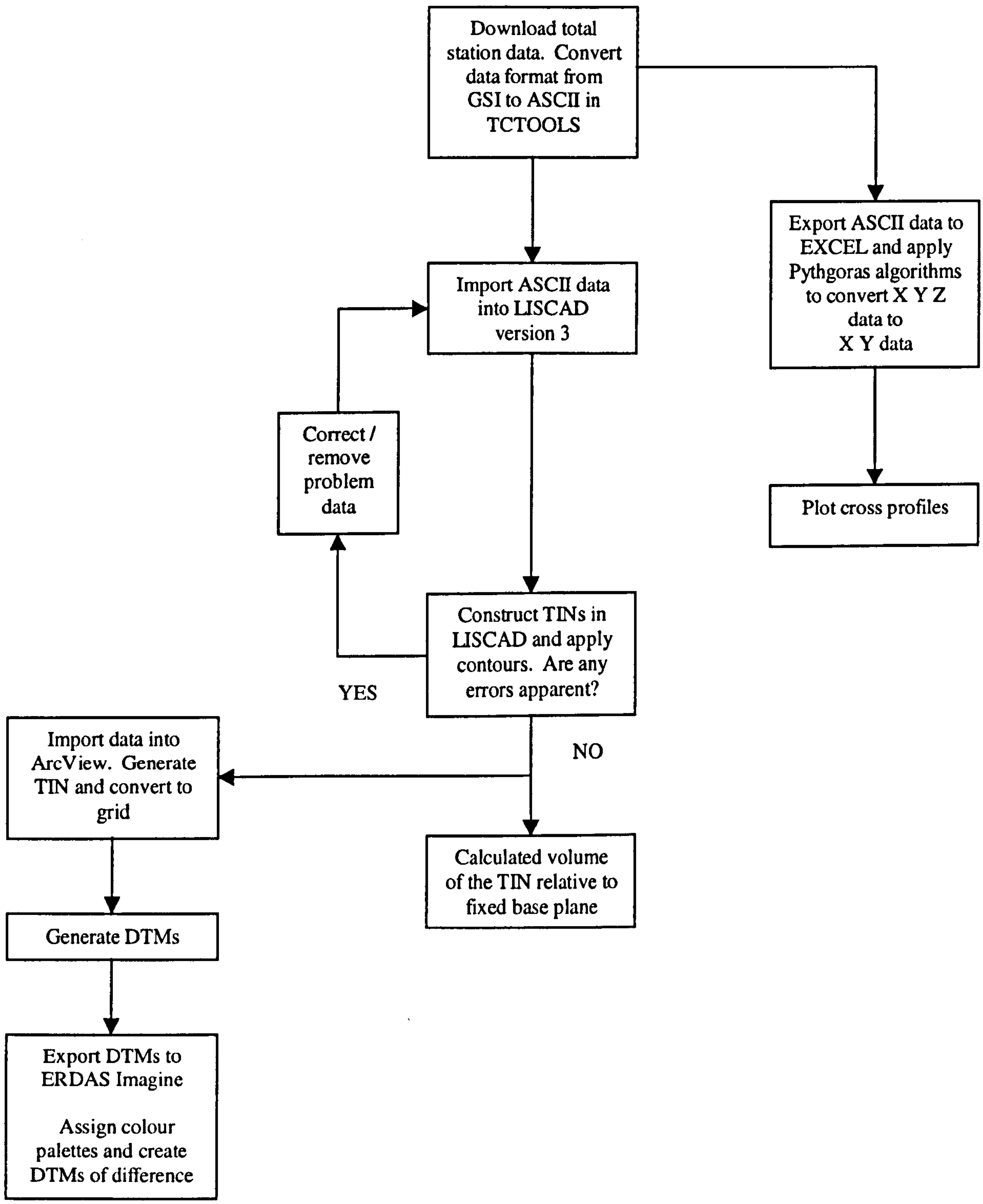
7.3.3 Digital Terrain Modelling

A schematic diagram of the procedures used to produce cross-profile outputs, surface interpolate the survey data and generate channel DTMs is presented in figure 7.3.

7.3.3.1 Data Preparation

Total station data were converted from the generic GSI to ASCII format in TCTOOLS version 6. Coordinate data were imported into LISCAD version 3. The X, Y and Z coordinate data were also loaded into EXCEL where Pythagorean-based algorithms were applied to convert the data to distance and height measurements for each section line. These sections could then be plotted.

Figure 7.3 Schematic diagram showing the procedures used to present and analyse field survey data.



7.3.3.2 Surface Interpolation

The data were tested for their ability to undergo surface interpolation using LISCAD version 3. Although only capable of a basic contour display the package enabled a fast method by which surfaces could be interpolated using a Triangular Irregular Network (TIN). Each of the TINs was defined using a maximum network search distance of 25 m. This prevented anomalous floodplain surfaces being generated as a result of the program triangulating section end points which were more than 25 m apart which had the effect of extending the convex hull (the outer boundary of the triangulated surface). The modelled vector data were visually checked for errors and anomalous survey points, which disrupted the model integrity, were either rectified or removed after checking the original data files. Channel volume for each channel TIN was calculated relative to a common base plane set at the elevation of the floodplain.

Surface interpolation was also undertaken in ArcView, again using a TIN, primarily to generate models from which morphologic change could be easily visualised and to allow conversion of the TIN models to grid format DTMs. The ArcView TIN output was converted to grid format, effectively converting the spatially variable vector data to spatially comparable raster data (of 0.5 m resolution) with each of the raster values determined from the TIN surface. Converted data sets were then exported to ERDAS Imagine in which DTMs of the channel were created. DTM outputs were differenced to highlight patterns of deposition and erosion and assigned specifically designed palettes to aid data visualisation.

7.4 Reach scale channel changes – results

7.4.1 Planform adjustment

Planform maps, drawn from field surveys, were generated for the CV and BH reaches. Planform maps have not been generated for the DD reach because it displayed lateral stability throughout the study period. Hence, planform maps drawn from field survey do not show any planform adjustment. Each planform map is accompanied by the daily mean stage value, recorded at Caersws and extrapolated upstream to Trefeglwys using the linear relationship presented in figure 5.7. The DMS value is provided as a comparative guide as to whether the plotted position of the surveyed planform components (particularly the position of the gravel bar / water interface) has been heavily influenced by stage. It should be noted that the same issues pertaining to the quantification of gravel bar areas from rectified aerial photography (discussed in section 6.12) exist here and, therefore, limit the information which can be gained from these planform maps alone. However, sub aerial planform features presented on these maps, particularly the location of active banks and the break of slope taken to represent bankfull location, have been located via accurate field survey methods and hence provide a reliable method of gaining information about locations and magnitudes of bank retreat.

Accompanying the planform maps is a further field map showing the location and type of bank protection employed at the two reaches (figure 7.4). A mixture of riparian tree planting, bouldering and gabion emplacement have been used to protect the two reaches with the protection being focussed about the outer banks of meander bends. Locations of bank protection failure are also shown.

7.4.1.1 Chapel view planform adjustment

The planform maps for the CV reach are presented in figure 7.5. The major points bars (PB1 and PB2) are labelled. It can be seen that DMS ranges between 0.21 m during survey 1 to 0.43 m during survey 6. However, the DMS for this last period is considerably greater than that of the other four survey periods where the DMS range is only 0.07 m. Consequently, the planform maps between surveys 1 and 5 can be considered broadly comparable, whilst the planform map of survey 6 underestimates the relative extent of gravel bar areas.

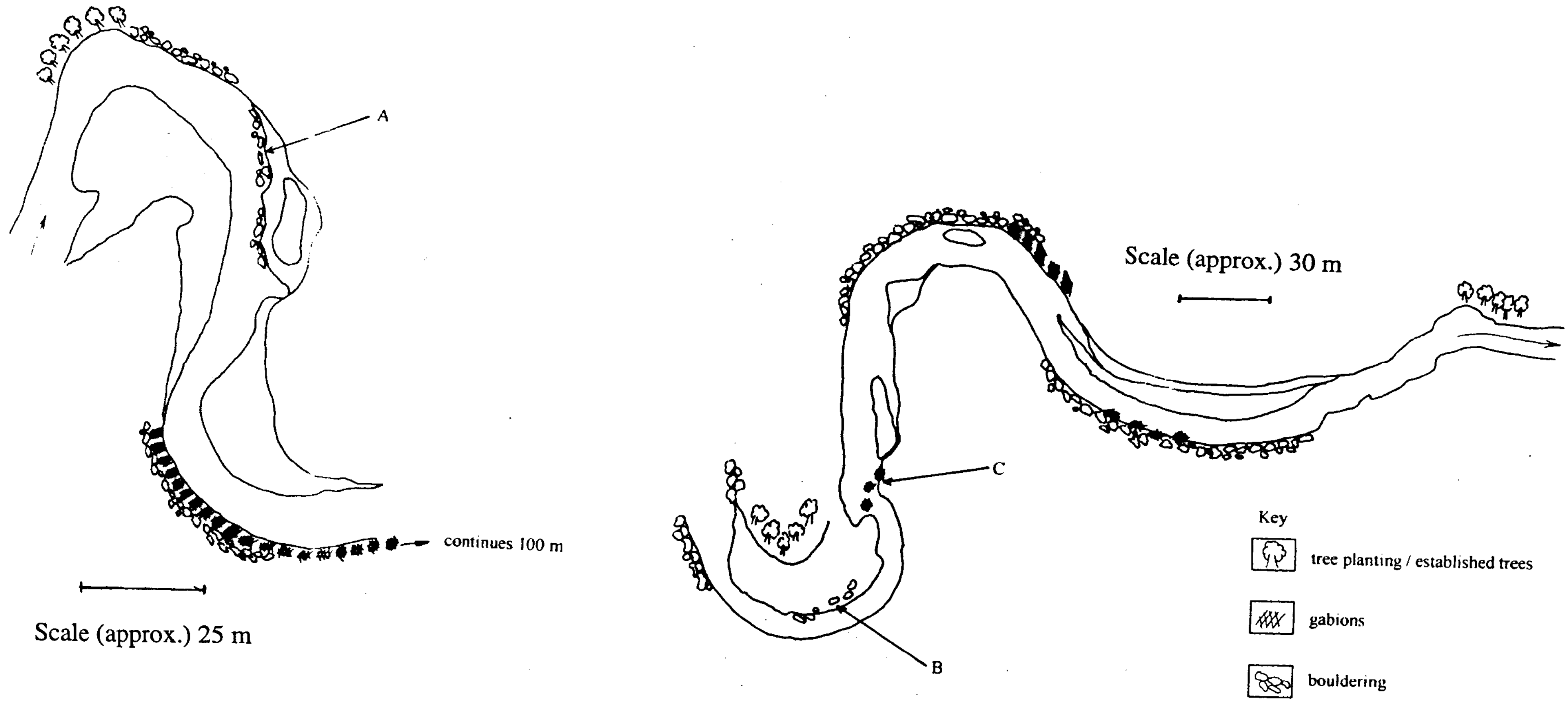


Figure 7.4 Location and type of bank protection in the Chapel View and Bodiach Hall study reaches. Locations A-C indicate areas where the protection has failed.

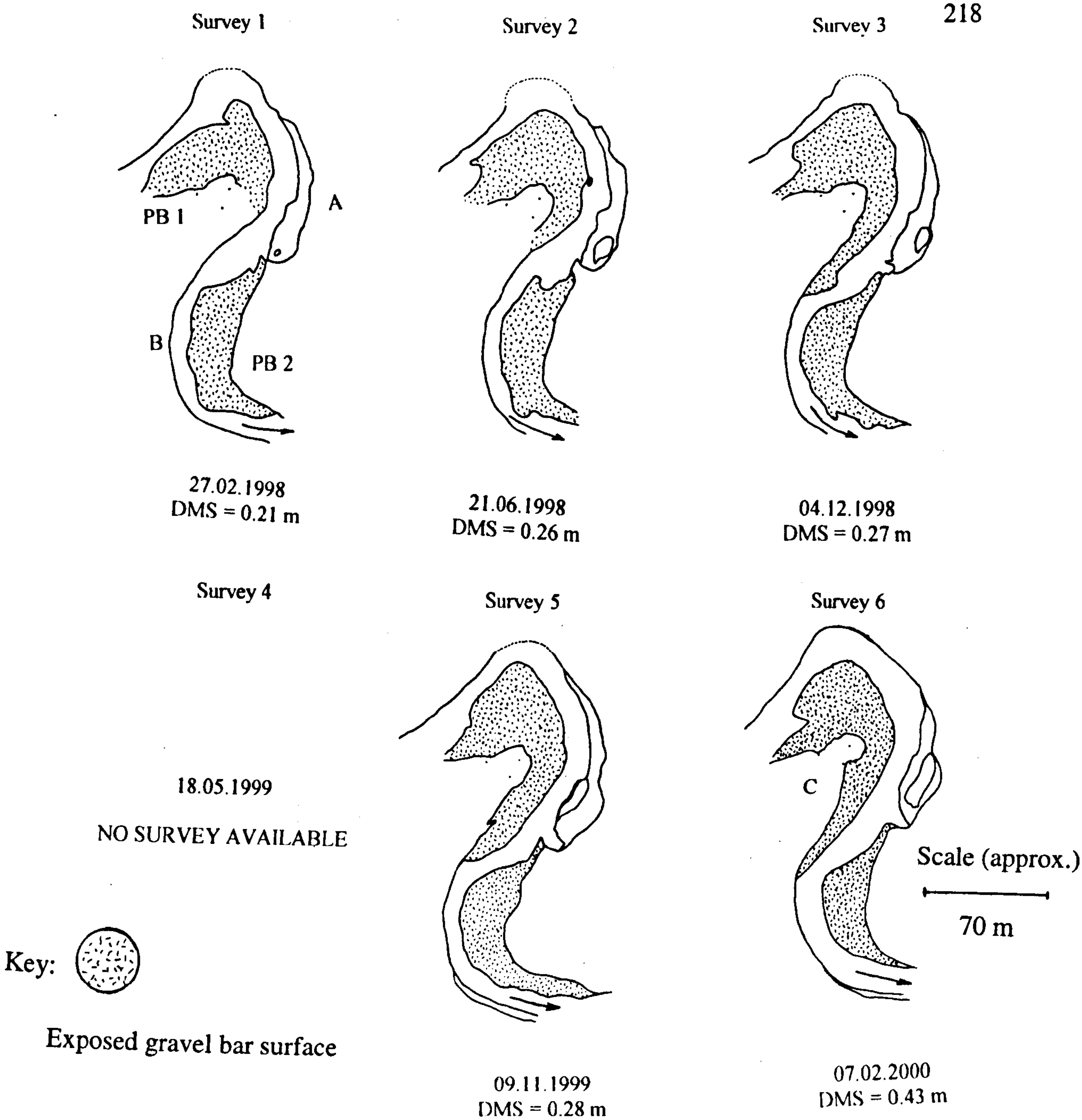


Figure 7.5 Planform maps of the CV study reach. Major point bars are labelled PB1 and PB2. A small pool in a normally inactive region of the channel is shown developing at A. B and C are further areas of erosion.

The immediately apparent feature of the planform sequence is the downstream extension of PB1 accompanied by southward channel migration and the removal of material from the upstream boundary of PB2. Erosion of the bank opposite to PB1 is almost totally restricted by extensive bouldering of the bank. Consequently, one can hypothesise that bed load deposition has been artificially forced downstream by high shear stresses maintained about the meander apex during high flows due to the inability of the outer bank to undergo erosion.

The deposition of gravel on PB1 has produced some erosion on the opposing bank (marked A on the planform map), but downstream of PB1 rather than directly opposite it. Here the bank protection (the inner bank line) has been breached and a water filled depression, isolated from the channel at low and moderate flows, has progressively increased in size accompanied by bank retreat. Refering to figure 7.2 one can see that cross-profile line 8 is located across the retreating bank. The cross-profile is presented in figure 7.6. The left bank is composed of composite material (i.e. composed of a mixture of coarse gravel and sand with a fine sand / silt matrix) and similar in fabric to those described by Winterbottom and Gilvear (2000). Cross profiles show it to have retreated 3.51 m between surveys 2 and 4 – equivalent to a mean time integrated rate of 3.9 m yr^{-1}. Also apparent from the plot is deposition on the left bank between surveys 4 and 6. This, however, is a product of the dumping of farmyard rubble into the channel by the farmer in an attempt to halt the bank retreat. The original bank protection boulders, placed in 1979, now form the central channel ridge.

Interestingly, gravel deposition, and erosion of the opposing bank (labelled B), appears to be significantly less marked at PB2. Some small-scale downstream extension of the bar (approximately 10 m) can be noted on the margin of the planform between surveys 1 and 2, however the barform area appears to extend little subsequently. Again, the outer bank at B is substantially engineered with protection provided by a mixture of boulders and gabions. Here though, the engineering has remained effective along the entire outer bank and continues downstream for a further 100 m. This engineering would therefore appear to indirectly restrict deposition on this bar, especially when one considers that composite bank erosion at A may well form a substantial local upstream source of bed load to PB2.

The final location of significant planform change is the inner bank of PB1 (labelled C). Although stable for the majority of the study period, it can be seen to retreat by up to 2 m between survey 5 and survey 6, however the retreat is very localised and compared to the bank retreat at A change is small.

7.4.1.2 Bodiach Hall planform adjustment

The planform maps covering the Bodiach Hall study reach are presented in figure 7.7. DMS varies between 0.17 m at survey 1 and 0.44 m at survey 5. As with the Chapel View planform maps this variation in stage means that the plan size of gravel bars may be underestimated and gently sloping features may show apparent retreat for planforms covering surveys 5 and 6 when compared to the earlier maps. A further obstacle to the analysis of the planform maps of Bodaich Hall is the discontinuous nature of the first four maps in the sequence (shown by features represented by a dashed line). Initially, only the channel surrounding point bars PB3 and PB4 were planform mapped and no planform is available at PB4 for survey 4. Consequently, the missing information has been filled in using data from the closest subsequent survey in the sequence. In the case of the planform maps of surveys 1 and 2, the channel between PB3 and PB4 is reconstructed from the maps of survey 3. This probably provides a reasonable approximation because the left bank of the channel is protected by a mixture of gabions and boulders, and hence is largely stable between PB3 and PB4. The channel reconstruction downstream of PB4 for maps 1 - 4 in the sequence is far less certain and probably contains significant error.

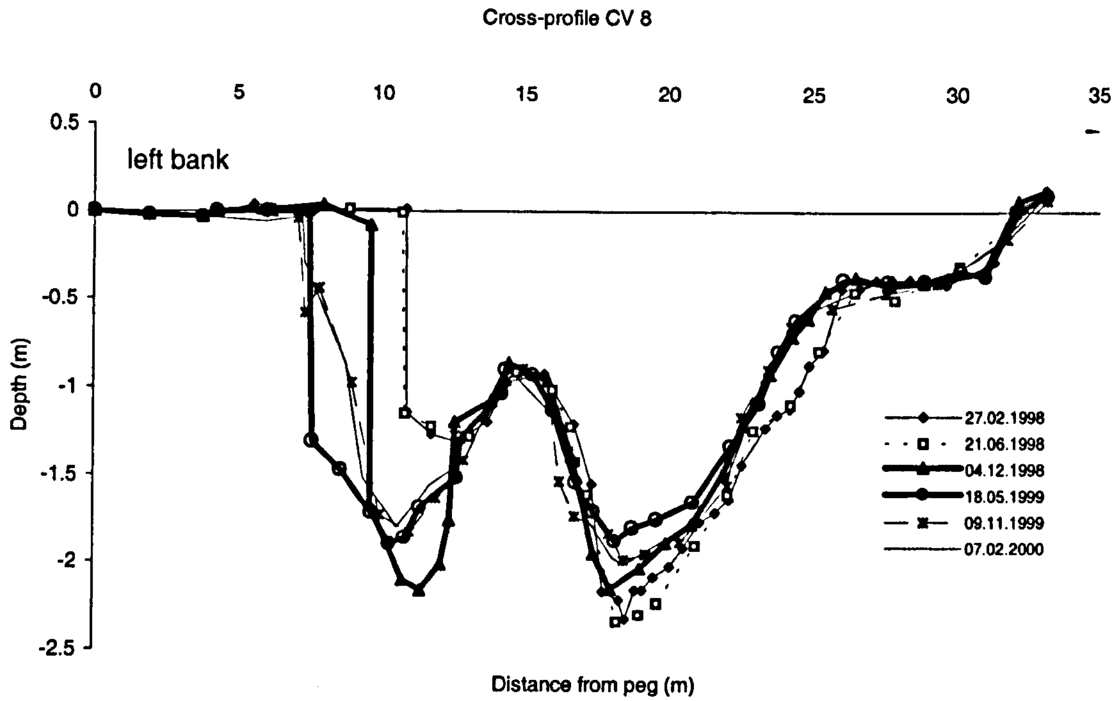


Figure 7.6 Channel morphology change at cross profile CV 8 throughout field surveys.

The most extensive zone of planform change which occurs in the Bodiach Hall reach can be linked to the development of point bar PB3. It can be seen that progressive deposition on the point bar is accompanied by erosion of the unprotected outer bank (labelled A) and the meander form responds by progressively becoming more asymmetrical. Cross-profiles BH7 and BH8 are located here and are presented in figure 7.8. The elevation of the gravel bar surface increases by up to 1.55 m between survey 1 and survey 4, and the bar can clearly be seen to migrate laterally. In order to maintain the channel size and hence accommodate flow, the outer bank is eroded. The rate of retreat can be seen to progressively decrease between survey 1 and survey 4 and then increase again between surveys 5 and 6. Between surveys 1 and 4 the maximum recorded mean time integrated bank retreat rate is 5.53 m yr^{-1}.

The planform change around PB4 is less pronounced. The bar can be seen to decrease in plan size between surveys 1 and 2 and subsequently extend in an upstream direction.

There is no noticeable bank erosion on the opposite bank at PB4. Again, this is due to bank protection in the form of gabions and boulders, which have remained effective since their emplacement in 1979. Therefore the potential for gravel bar migration in a cross stream direction is impaired.

The planform maps suggest some morphologic change between PB3 and PB4 in that the numerous small lateral and unit bars in this zone experience apparent growth and retreat between surveys 3 and 6. This zone is a riffle unit and hence has low water depth. The planform change is therefore most likely a result of variability in the DMS rather than significant morphologic change.

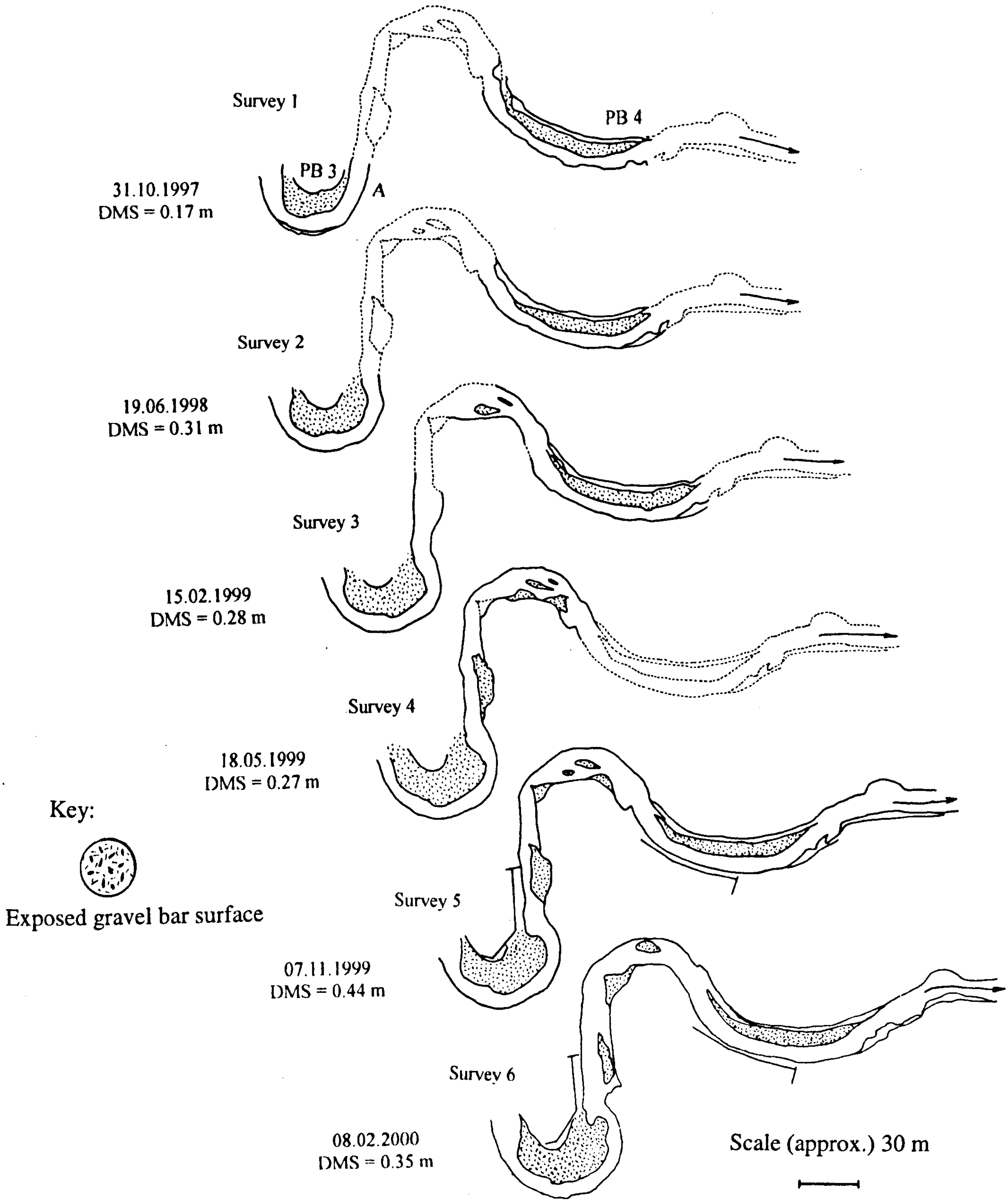
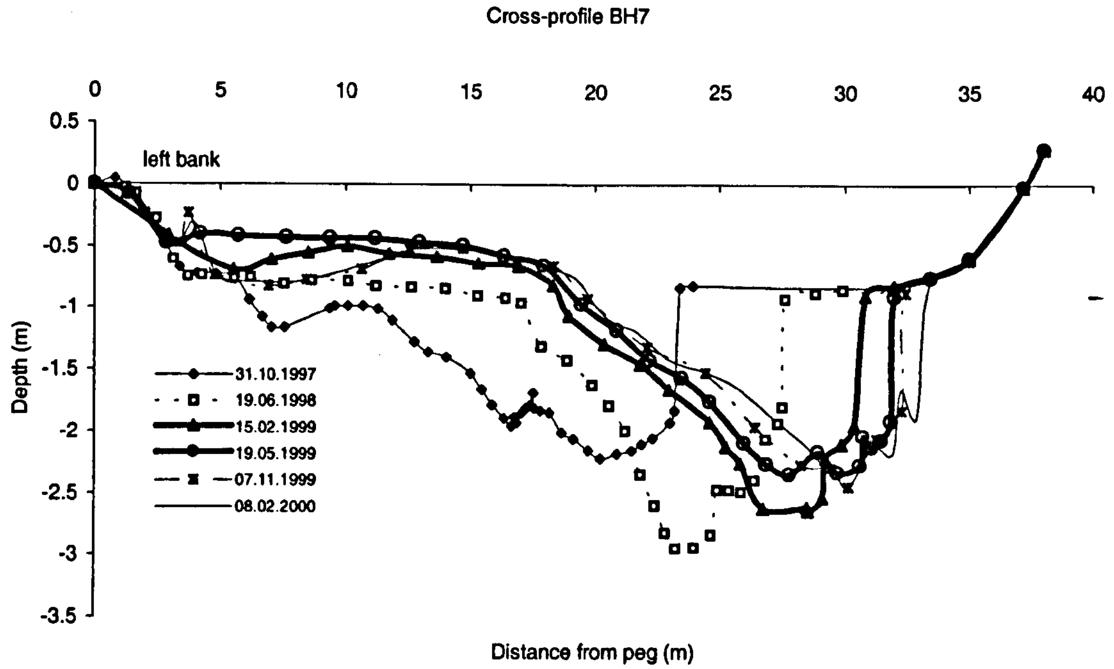


Figure 7.7 Planform maps of the BH study reach. Major point bars are labelled PB3 and PB4. The major region of bank erosion is labelled A.



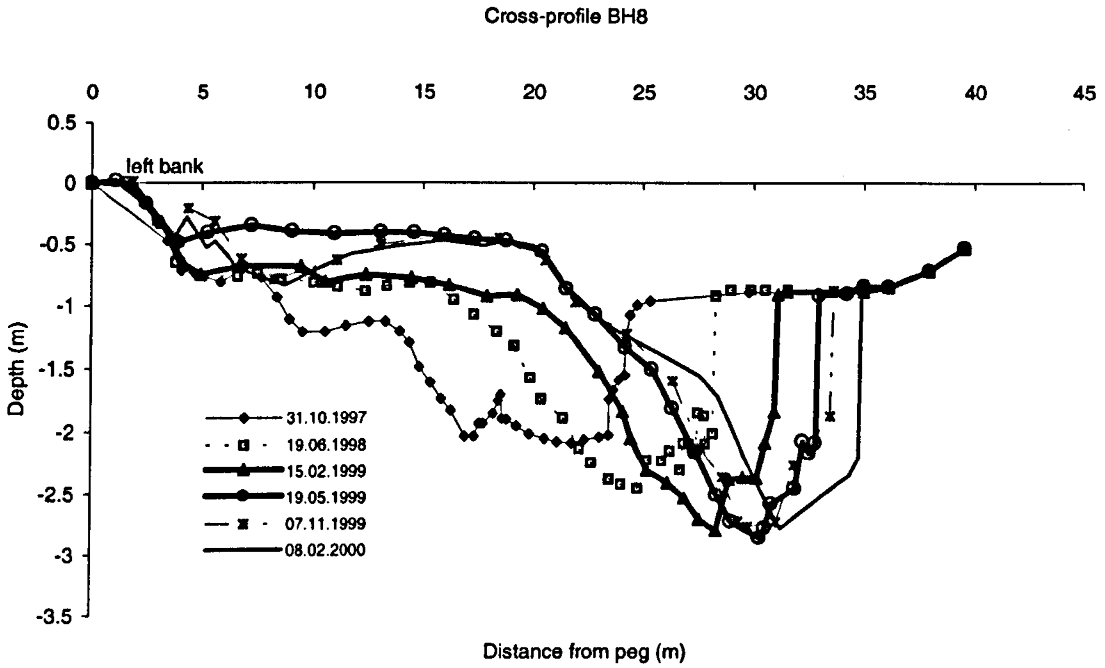


Figure 7.8 Channel morphology change at cross profiles BH 7 and BH 8 thoughout field surveys.

7.4.1.3 Planform change – summary

The planform maps presented here are able to demonstrate the positions and, in consultation with relevant cross-profiles, approximate magnitudes of the major zones of lateral channel change. Both the CV and BH reach exhibit localised zones of highly active planform where mean rates of bank retreat can be in excess of 0.015 m day^{-1}. These zones are seen to occur at, or close to, meander bends where significant, apparently expanding, stores of gravel exist as point bars, and this is perhaps unsurprising given the sinuous nature of the study reaches. The planform maps are also able to infer a difference in the development of large point bars between zones where the opposite bank is engineered and zones where bank engineering is absent. Where bank protection is absent or fully breached (PB3) high rates of lateral change are seen and the bar growth occurs in a mainly cross stream direction. At PB1 and PB4, where the opposite bank is protected, gravel bar development can be seen to occur either in an upstream or downstream direction, and in the case of PB1 opposite bank erosion occurs where the bank protection measures have become compromised. Moreover, where bank protection exists over a considerable distance from the barform, its development is very minimal (PB2). This suggests that river engineering of the Trannon, although ineffective in many cases (Leeks et al. 1988) may still have a large impact on the restriction of local planform development.

7.4.2 Reach-scale volumetric response – analysis of reach scale TINs and DTMs

Channel TINs and DTMs have a significant advantage over planform maps in that they are able to highlight both areas of lateral erosion and deposition and erosion and deposition in the vertical plane. The information contained is also not restricted by water in the channel, being constructed from survey points which have been measured through the water. The construction of TINs for data sets containing information in the X, Y and Z directions means that each of the Delaunay triangles of the network (the triangles connecting the network of survey points with the specific property that their internal angles are as nearly equiangular as possible) can be raised and tilted to form a plane. The collection of all of the triangles in the network forms a representation of the surface from which a volume, relative to a fixed plane, can be obtained. The accuracy of the TIN surface representation is primarily a function of the density of the survey network. If an extremely high level of detail is required a dense survey network is also necessary.

A restriction of TINs becomes apparent when one wishes to volumetrically compare triangulated land surfaces covering the same area, but at different dates. If the TIN for each survey iteration has been constructed from the same number of points gathered at identical locations the Delaunay triangles are fully comparable in terms of dimension, and may be subtracted from one another to quantify change. However, where survey iterations contain survey points gathered at spatially variable locations, the Delaunay triangle dimensions for each surface representation become variable and hence, cannot be subtracted from one another. In such circumstances it is necessary to convert the TIN to a grid, in which the surface is represented by a rectangular mesh of points joined with lines, creating a grid of uniformly sized square cells which represent the surface. A land surface representation in the grid domain is called a digital terrain model (DTM). If the resolution of the grid cells is the same for all survey iterations, one can simply subtract one DTM from another to identify and quantify areas and volumes of surface change. Necessarily, the conversion of a TIN to a GRID requires X, Y, Z values for points on the ground where the X, Y, Z have not been measured. It extracts these values directly from the linear triangulated surfaces of the TIN, thereby assuming that the TIN provides an adequate representation of the surface.

Here, all channel volumes have been calculated directly from the channel TINs relative to a common base plane and convex hull applied throughout the channel surveys. The channel DTMs and DTMs of difference have been generated from by linear resampling of the original Delaunay triangle TINs (Pueker et al., 1978; Brasington et al., 2000), which have been subsequently converted to a 0.5 m grid.

7.4.2.1 Reach-scale change – Bodiach Hall

Channel DTMs are shown in figure 7.9. The DTMs at BH have been constructed from only the section data – planform survey data has not been included to ensure that the Delaunay triangles form in a downstream rather than cross stream direction. The left hand peg elevation at each section has been standardised to 0 m – this a reasonable approximation for most sections because the peg is located on the valley flat (the main, inactive floodplain surface) which has a very low downstream gradient of < 0.004. However, the model does break down slightly (in the form of a ridge across the bed) at section BH 40, where the peg is actually located approximately 1 m below the valley flat. The ridge, though, appears on all images and hence it's effect is removed when DTMs are differenced. Contours have been set as soft break lines meaning a smooth, linear interpolation between contour line locations is achieved. Colours indicate elevation at 0.5 m intervals according to the given key. It should be noted that the DTMs representing the channel at surveys 1 and 2 have had the channel between BH 10 and BH 20 filled in from survey 3. Consequently, no change occurs in this zone in the early DTMs.

The DTMs appear to represent the channel morphology well at Bodiach Hall. The low elevation channel line is clearly visible as are the higher elevation barforms. Indeed, meso-scale bedforms seem well defined, in particular the riffle and pool units which occur between sections BH 10 and BH 20 and the point bars PB3 and

PB4. The curve of the channel margin is, in the main well represented, particularly in the higher resolution survey zones between BH 1 and BH9 and BH 20 to BH 47. The lower section resolution between BH 10 and BH 20 has introduced some angularity into the channel margin. However, the clearly defined pool and riffle units would suggest that the decrease in section resolution has not had a serious impact on the ability of the model to represent meso-scale bed form in this region of relatively gentle downstream variability in bed morpholgy.

The DTM series confirms much of the channel change inferred from the planform mapping. Clear evidence of large-scale deposition and outer bank erosion is seen at PB3, and extensive barform development at PB4 is also visible, particularly between surveys 1 and 2. However, the DTM is better able to display the extent of lower magnitude changes. A good example is provided by the small, low relief, right bank lateral bar immediately downstream of PB3 (labelled LAT1 on the DTM sequence). The planform maps (figure 7.7) show its plan size to be decreasing slightly between survey 4 and 6, and the feature does not even appear on the planform map for survey 3. Inspection of the DTMs shows the erosion of this feature to be more pronounced than the planform maps suggest. Indeed, by survey 6 the bed elevation at the bar location is almost the same as that of the main channel area indicating that the bar has actually been almost totally removed. This may be in response to the lower yields of bed load sized material to the channel from bank erosion at PB3 as the lateral migration rate slows after survey 4.

The DTM also has the advantage of displaying the morphological response of submerged parts of the channel. Of particular interest is the different channel response, in terms of depth, at PB3 and PB4. Bar development at PB4 does not cause associated outer bank erosion because of the stable, engineered bank configuration. Consequently, the channel responds by deepening which can be seen by the introduction of pink and blues between surveys 3 and 6. In contrast, the depth of the channel at PB3 changes little as it migrates except for a localised increase in depth of about 0.5 m between surveys 1 and 2.

DTM difference models between each consecutive survey period are presented in figure 7.10. The difference models show the product of the subtraction of later models from earlier. Areas of deposition have been assigned increasingly dark grayscale shades, with the darkness being directly proportional to the magnitude of the change. Conversely, areas of erosion have been assigned increasingly light grayscale values proportional to the magnitude of change. Zero change has been assigned the background shade, which should be used as the reference shade in interpreting the figures. Difference models 2-1 and 3-2 show no change between sections BH 10 and BH 20 due to no survey data being available in this region for these periods. Where the channel margins are defined by vertical, stable banks there are occasional thin strips of very high magnitude elevation change indicated by thin stripes of black or white which are visible. These stripes are probably indications of survey error, where tilt of the survey pole has produced a false cut or fill of the bank which shows up in the difference models as erosion or deposition. Commonly they will exist on several of the difference DTMs and alternate between cut and fill with successive surveys. A good example is visible on the near-vertical stable bank opposite PB4.

The channel in the vicinity of PB3 and PB4 is again shown to contain the most active region of morphologic change. The dark shades representing deposition of gravel onto PB3 are easily seen together with the associated outer bank erosion shown up white. In the same way, deposition of gravel onto PB4 is clearly displayed in DTM 2-1. The difference DTMs also highlight a region of change immediately downstream of PB4 which has not yet previously been emphasised (figure 7.10b). Here survey 5-4 shows substantial deposition of gravel across almost the entire channel width, whilst survey 6-5 shows its subsequent removal. This morphologic change can be linked to the filling and scour of a deep pool, which has remained relatively unchanged on previous surveys. Interestingly, following deposition of bed material into the pool, the channel immediately downstream experiences increased instability. In particular, the white shades indicate mid-channel degradation, probably in response to the restricted bed load supply from immediately upstream.

Between section BH 10 and BH 20 the channel remains relatively stable throughout the available surveys. The upstream most pool can be seen to aggrade slightly in DTM 4-3, followed by degradation in DTM 5-4. Also visible is a region of inner bank erosion between surveys 5 and 6. However change in this section of the channel appears of substantially lesser magnitude than the change around PB3 and PB4.

Total volume change of the reach is presented in table 7.2 and figure 7.11. Despite substantial local high magnitude morphologic change the entire reach volume change is of lower magnitude. Overall the channel experiences cumulative degradation equivalent to 3.4 % of the original reach volume. However, the results are influenced by the lack of surveys between BH10 and BH 20 for survey periods 1 and 2. This can be seen by the substantially lower volume changes for these periods and the fact that they record reach aggradation – unsurprising given the barform development through these periods. Consequently, it is possible that the reach has degraded slightly more than the cumulative figure presented here.

Table 7.2 Channel volume at the Bodaich Hall study reach 30.10.1997 – 06.02.2000

Survey number	Date	Reach volume (m^3)	Volume change (m^3)
1	30.10.1997	9893.1	
2	19.06.1998	9874.7	- 18.4
3	15.02.1999	9849.7	- 25.0
4	19.05.1999	9959.4	+ 109.7
5	07.11.1999	10045.2	+ 85.8
6	08.02.2000	10231.7	+ 186.5

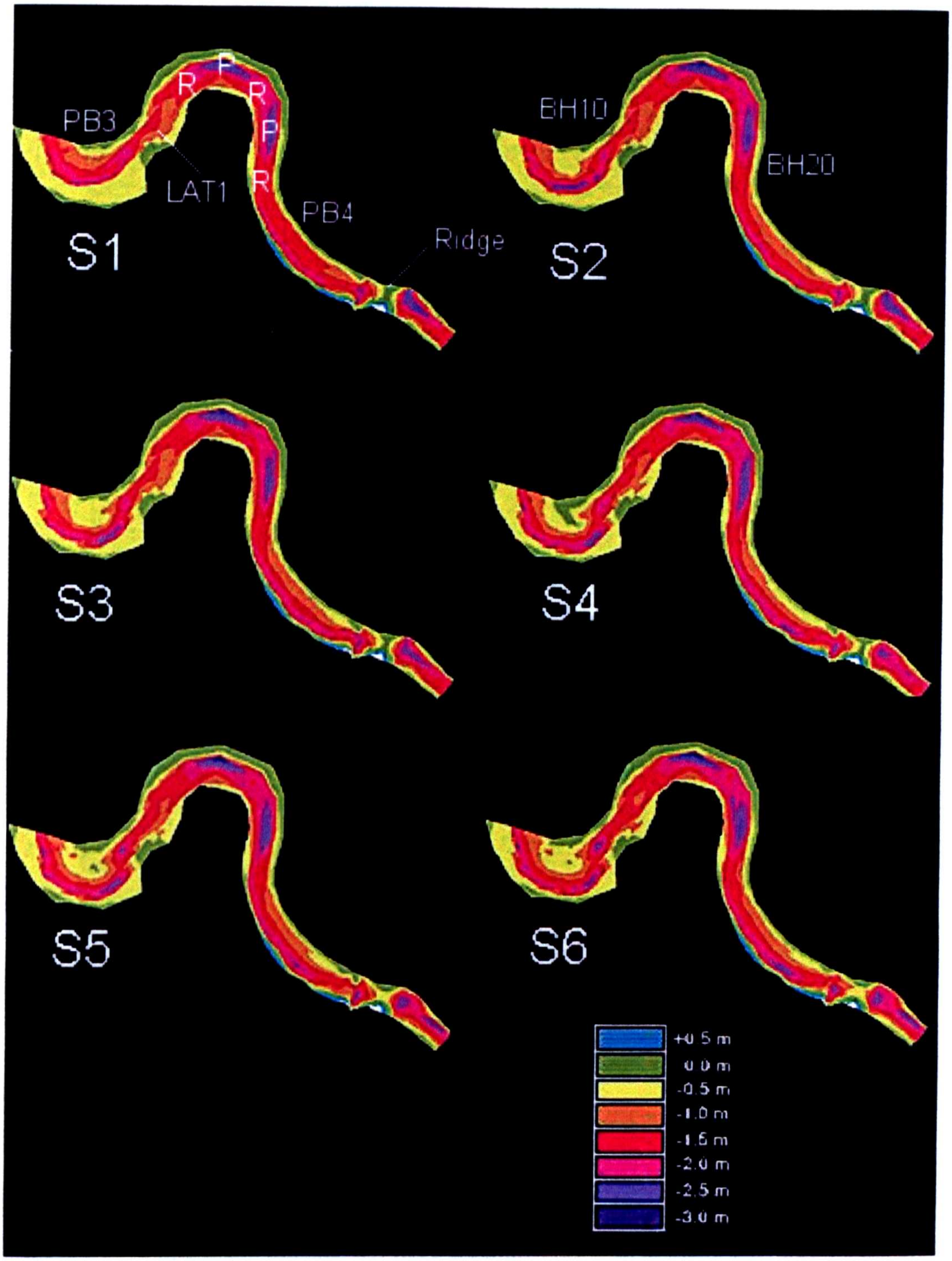


Figure 7.9 Channel DTMs of the BH reach. Scale is approx. 1:3500.

Survey numbers are indicated S1 to S6. Also shown are locations of locations of point bars PB3 and PB4, pools (P) and riffles (R), location of lateral bar (LAT1) and the anomalous model ridge. Flow is left to right. **Note**: region between section lines BH10 and BH20 in DTM S1 and S2 is interpolated from data collected during S3 and is not necessarily representative of channel morphology.

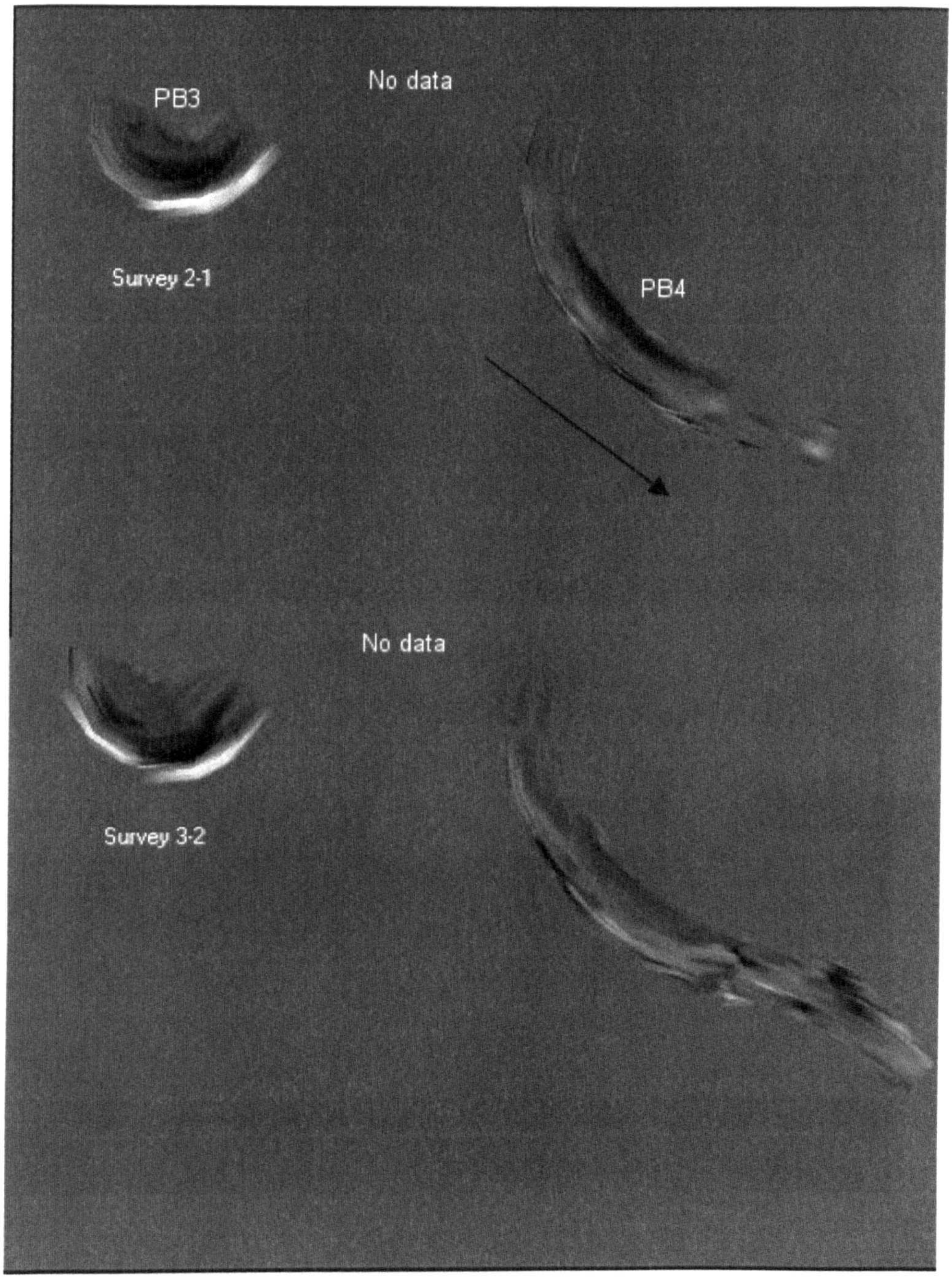


Figure 7.10 (a) Differenced DTMs of the BH reach for surveys 1-3 (scale approx 1:1850). Shades lighter than the background represent erosion and shades darker than the background represent deposition. Arrow indicates flow direction.

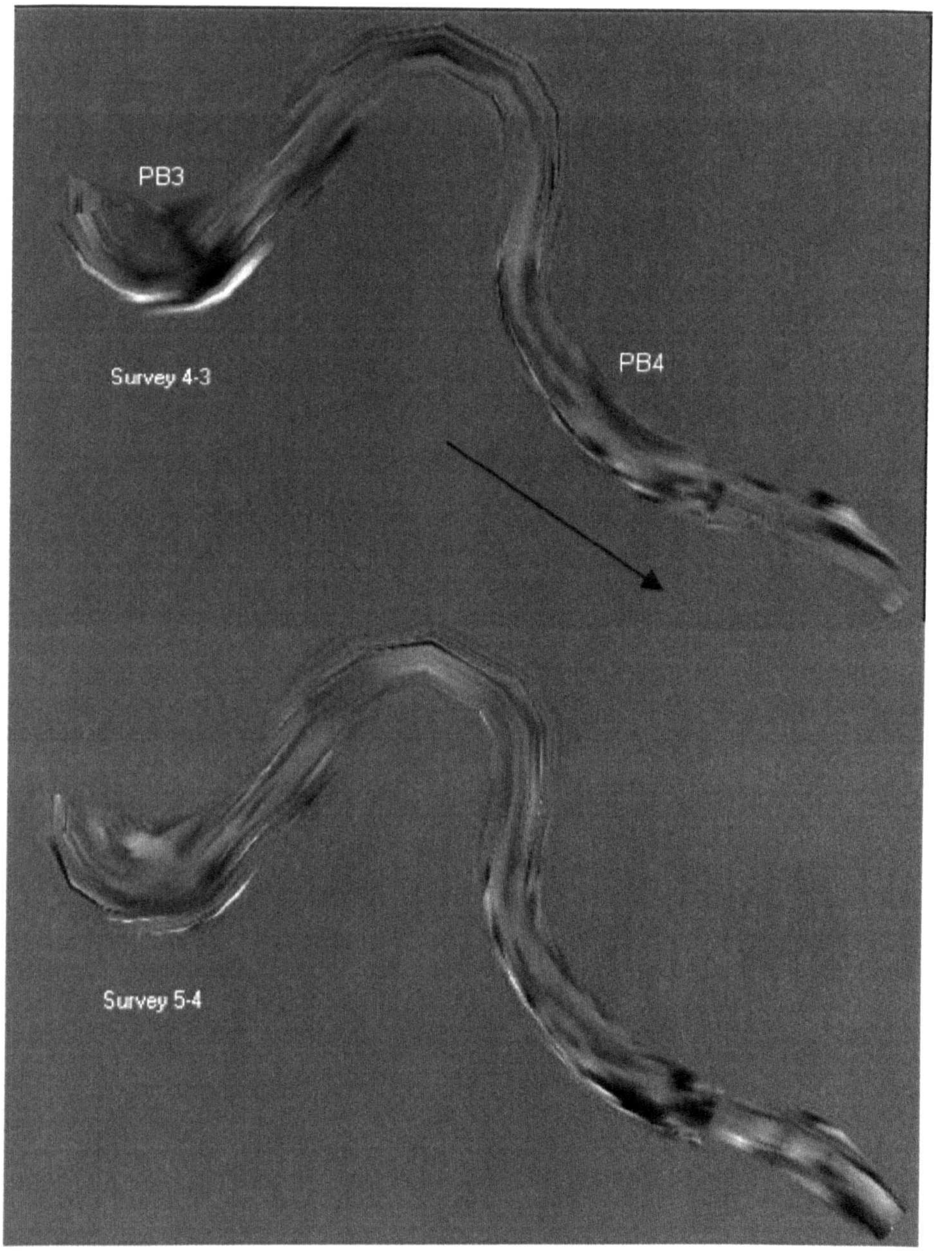


Figure 7.10 (b) Differenced DTMs of the BH reach for surveys 3-5 (scale approx 1:1850). Shades lighter than the background represent erosion and shades darker than the background represent deposition. Arrow indicates flow direction.

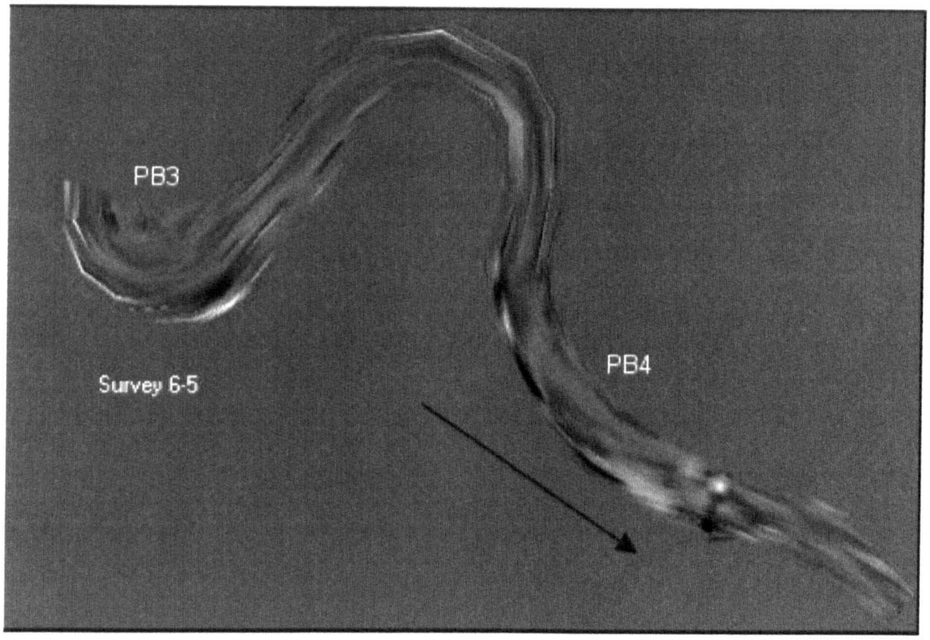


Figure 7.10 (c) Differenced DTMs of the BH reach for surveys 5-6 (scale approx 1:1850). Shades lighter than the background represent erosion and shades darker than the background represent deposition. Arrow indicates flow direction.

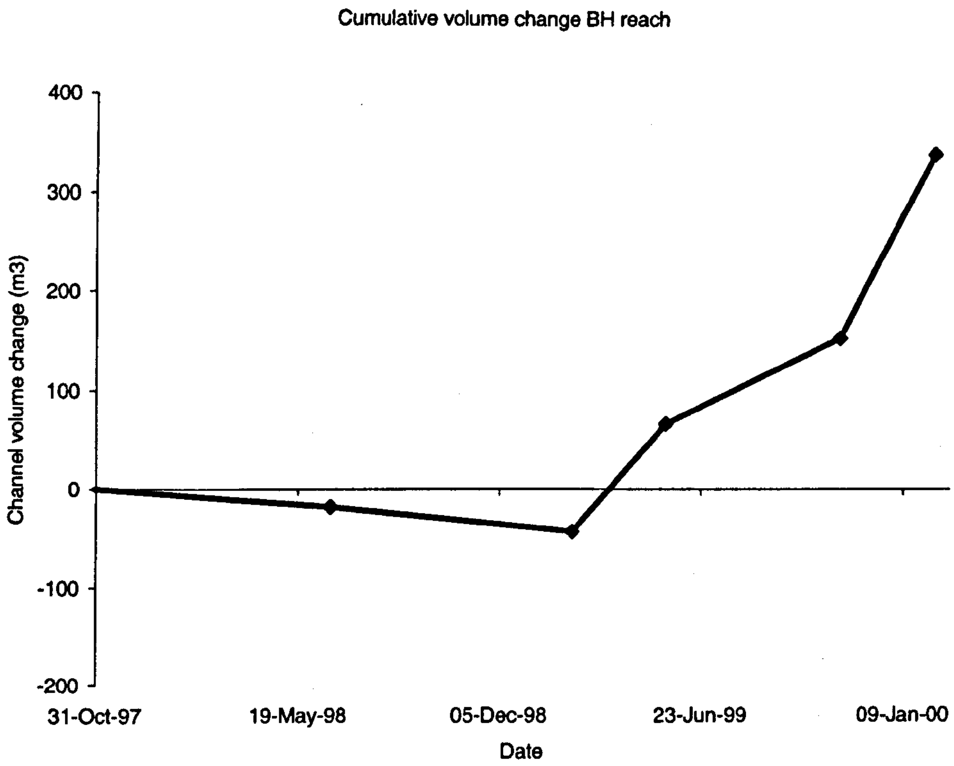


Figure 7.11 Cumulative volume change in the BH reach. Positive values show degradation.

7.4.2.2 Reach-scale change – Chapel View

The Chapel View DTM sequence is presented in figure 7.12. In contrast to the BH DTMs the planform surveys are also included in the data set from which the DTMs are built. This has been done in an attempt to improve the quality of the models which have been limited due to the higher section spacing employed at this reach. Inclusion of planform data has meant that standardisation of all section peg elevations to zero meters has not been possible and the valley flat elevation can be seen to decrease in a downstream direction.

The quality of the CV DTMs is impaired relative to the BH models. The main channel line is easily visible on all models, however the channel margin, particularly opposite PB2 during survey 6, can be seen to be highly irregular. This is in contrast to the planform maps. Indeed, field examination of the engineered bank opposite PB2 shows it to be sub-vertical and a morphologically regular feature. The spatial integrity of the model is also called into question when one examines the main channel, which appears to possess numerous small, isolated pools. Comparison of the DTM with the section line positions shows these 'pools' are commonly located close to a section line inferring that the original TIN from which the DTMs have been created has been constructed in such a way that the Delaunay triangles are forming preferentially in a cross-stream direction, meaning that the interpolation of channel elevation data is discontinuous in a downstream direction. The network of triangles making up the surface is particularly coarse, and in several locations triangular patterns in elevation change can be seen.

A further model error can be seen in survey 1. A line of high elevation (coloured yellow) crosses the channel downstream of PB1 in the position of section line CV 9. Subsequent checking of the data set has revealed this line to be a result of input data error, in which variability in elevation across the section has been artificially restricted. The error does not exist on subsequent surveys.

The extent to which meso-scale bed form change can be analysed is restricted by the questionable surface representation provided by the TIN. Whilst PB1 can be seen to migrate northwards through surveys 1-3, the downstream migration, shown by the planform mapping to be a more significant feature, is not well shown. Elevation can be seen to increase at the right channel margin downstream of PB1 between surveys 3 and 4, akin to the barform development shown on the planform maps, but the elevation can be seen to decrease again in subsequent surveys. Barform morphology at PB2 would appear somewhat better represented by the model. The bar can be seen to prograde between surveys 1 and 2 and subsequently remain fairly stable. However, the southward channel migration, clearly shown from the planform mapping, is not as clear in the DTMs. The pink colours, representing the deepest regions of the channel, show some southward migration, especially if surveys 2 and 4 are compared, but the discontinuous nature of the downstream surface representation makes visual analysis difficult.

The differenced DTMs at the CV reach show deposition as dark shades and erosion as light shades. The differenced DTMs are presented in figure 7.13. The ability to extract detailed change information from the differenced DTMs for the CV reach is hampered by the poor model representation. The triangular shapes of the TIN can be seen clearly in the models, especially about the channel margin. This suggests that the modelled channel morphology is a function of the TIN construction rather than the TIN construction being a good match of the channel morphology.

PB1 shows some evidence of prograding, particularly in the period between surveys 3 and 4 where a dark area is clearly visible at the downstream margin of the bar. This region however appears to experience erosion in the following period which is in contrast to the planform maps. Opposite this region of deposition, bank erosion of the opposite bank is clearly visible. This is the location of the cross profile CV 8 and erosion at this site can be seen in surveys DTM 3-2, 4-3 and 6-5. Deposition at this location is also visible in DTM 5-4 and this coincides with the dumping of rubble in front of the bank previously mentioned. These features are encouraging in that they demonstrate that some changes detected in the sections and planform maps are indeed clear within the DTMs.

PB2 appears to change little throughout the surveys, despite contradictory evidence from the planform maps. Certainly, the large-scale bar development seen in the BH reach is not apparent here.

Reach volume estimates for the CV reach have been acquired from TINs fitted to the section data only. This allowed the section peg elevations to be standardised to 0 m, and hence allowed volume calculations to be made relative to a fixed, flat base plane as required by LISCAD without including air space above the level of the valley flat. The inadequacies in the survey resolution previously demonstrated mean that the channel volume estimates probably contain higher error than those for the BH reach. The volumes and volume changes are presented in table 7.3 and figure 7.14. In contrast to the BH reach, overall reach aggradation is estimated, however the cumulative total of 150.5 m^3 of deposition (4.33 % of the volume for survey 1) is probably of a small enough magnitude to be a result of the TIN inadequacies rather than representative of real change.

Table 7.3 Channel volume at the Chapel View study reach 30.10.1997 – 06.02.2000

Survey number	Date	Reach volume (m^3)	Volume change (m^3)
1	27.02.1998	3474.6	
2	21.06.1998	3384.2	- 90.4
3	04.12.1998	3432.7	+ 48.5
4	18.05.1998	3378.0	- 54.7
5	09.11.1999	3329.3	- 48.7
6	07.02.2000	3324.1	- 5.2

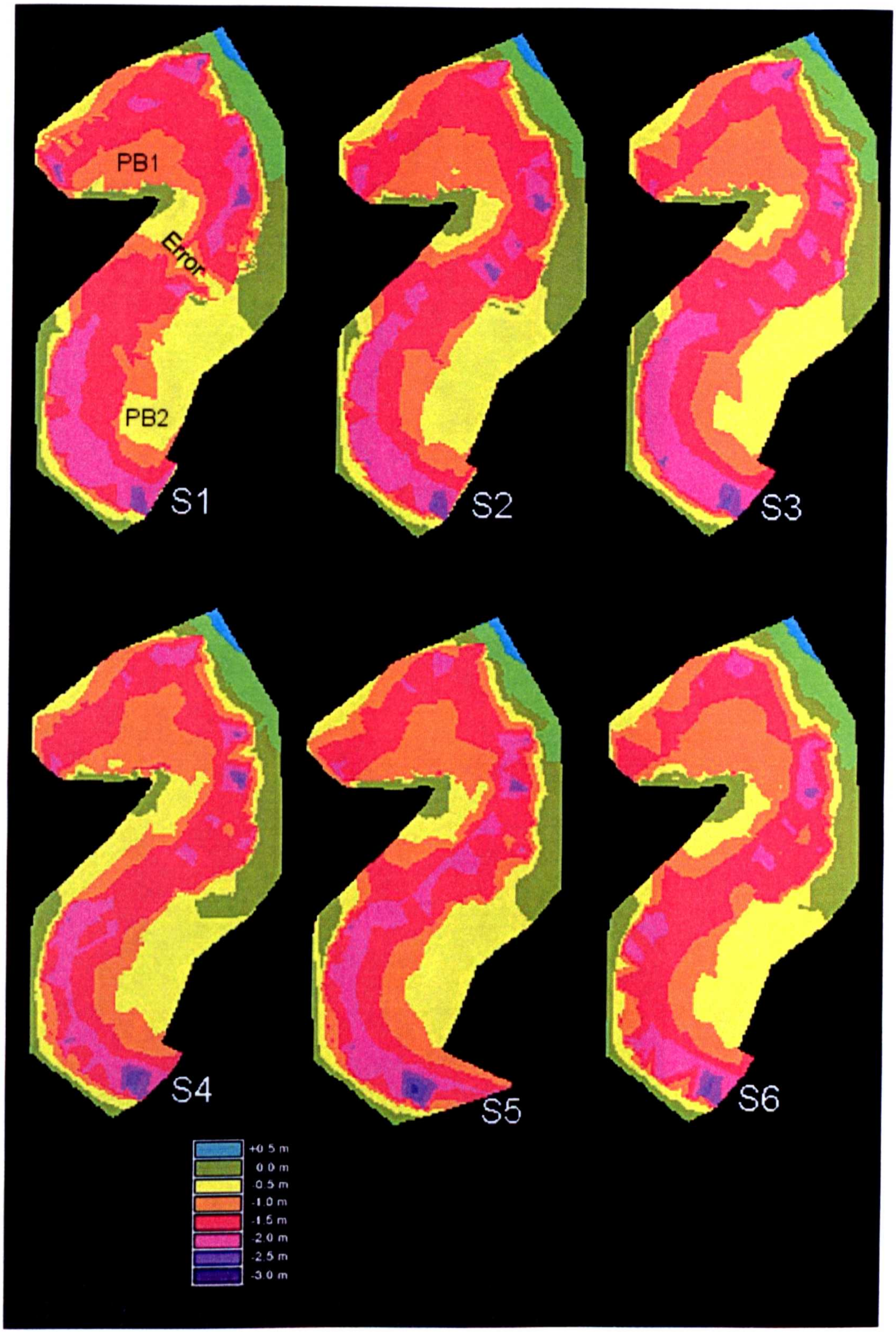


Figure 7.12 Channel DTMs of the CV reach. Scale is approx. 1:1250. Survey numbers are indicated S1 to S6. Flow direction is top to bottom.

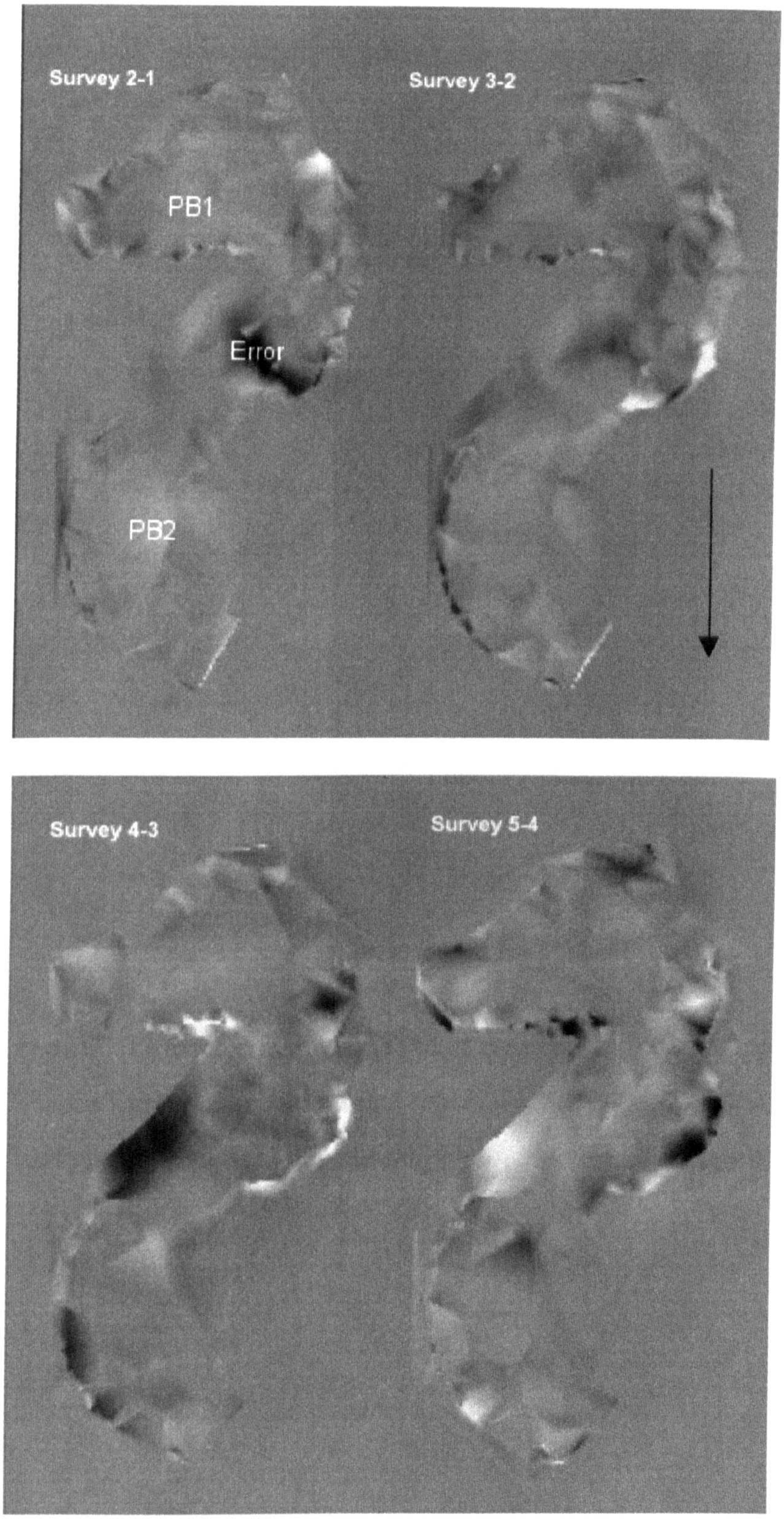


Figure 7.13 (a) Differenced DTMs of the CV reach (scale approx 1:1200). Shades lighter than the background represent erosion and shades darker than the background represent deposition. Arrow indicates flow direction

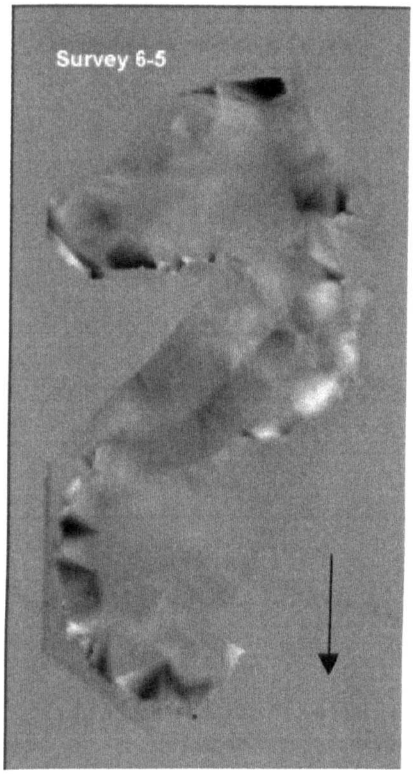


Figure 7.13 (b) Differenced DTM of the CV reach (scale approx 1:1200). Shades lighter than background represent erosion and shades darker than the background represent deposition. Flow is indicated by arrow.

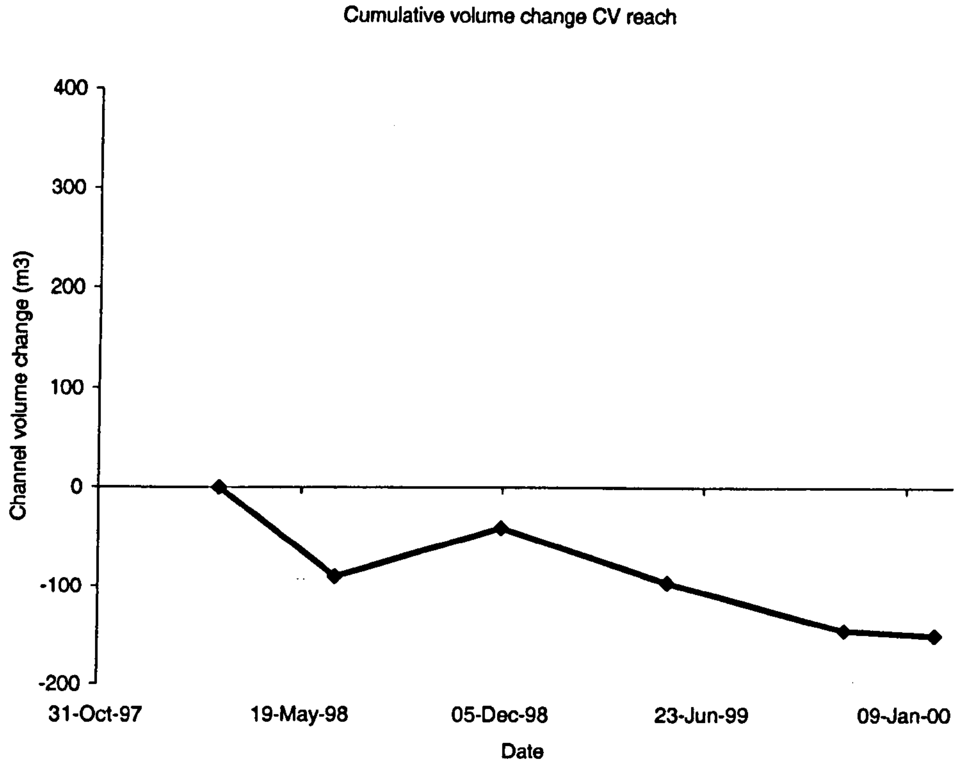


Figure 7.14 Cumulative volume change in the CV reach. Positive values show degradation.

7.4.2.3 Reach-scale change – Ddranen Ddu

The DTM sequence for the Ddranen Ddu reach is given in figure 7.15. The TIN from which the DTMs have been built is based on the cross profile data only. Peg elevations have been standardised to 0 m.

The reach DTMs show a generally stable channel, both in terms of lateral movement and changes in channel morphology. The banks contain two main benches, shown as the continuous light blue and green stripes running along the entire length of the models. There is some evidence of bed elevation change throughout the sequence, particularly towards the upstream end of the reach. Initial aggradation between surveys 1 and 3 is followed by degradation between surveys 3 and 5. The surveys also show some small-scale change in the position of the benches, however, compared to the lateral channel change at BH these movements are very minor.

Ddranen Ddu differenced DTMs (figure 7.16) conform to the format employed at CV, where erosion is shown as dark shades and deposition is shown as light shades. The differenced DTMs confirm a stable bed configuration, with shades consistently similar to the background throughout the sequence. It does, though, show in greater detail the changes in the position of the benches, which can be seen to undergo both erosion and accretion. Whilst the erosion of banks is a common fluvial dynamic, the channel-ward growth of banks is conceptually more difficult to reconcile. However, there are two possible explanations for the changes shown:

1. Rotational slumping of steep banks: The silt / clay banks at Ddranen Ddu fail by rotational slumping rather than the cantilever failure mechanism witnessed at BH. Such failure provides a mechanism by which the bank face can move channel-ward accompanied by a decrease in bank height. Slumping between surveys would therefore be represented in a DTM difference model as a band of dark shades at the front of the slump where the bank face has moved channel

ward and a corresponding band of lighter shades behind where the bank elevation has decreased.

2. Tilt of the survey pole: Pole tilt when surveying the break of slopes associated with the near-vertical bench faces could cause positional error, mainly in the X and Y directions, of the survey point. If the pole was tilted towards the channel the bank would be seen to undergo accretion. However, there would be no zone of reduced elevation behind the feature. The feature would also be discontinuous in a downstream direction.

A good example of channel-ward bank movement is provided in DTM 3-2, where a continuous dark band (labelled AB) can be seen running along almost the entire length of the DTM on the right bank. Behind the dark band a large proportion of the DTM is shaded lighter than the background. This pattern conforms to that expected as a result of rotational slump. However, there are examples where channel-ward bank movement is not accompanied by a decrease in elevation behind the bank face. One such example can be seen on DTM 4-3 (labelled CD). Here, pole tilt is more likely the cause of the apparent bank migration. The impact of pole tilt (which is a random error when considered in terms of the entire survey, causing both false bank retreat and accretion) is likely to be limited on the overall channel volume calculation.

Channel volumes for the Dranaen Ddu study reach are presented in table 7.4 and figure 7.17. Channel volume remains extremely stable throughout the study, with slight degradation occurring between surveys 4 and 5. In total the channel degradation is 16.9 m^3, equivalent to an increase of just 2.5 % between surveys 1 and 5.

Table 7.4 Channel volume at the Ddranen Ddu study reach

Survey number	Date	Reach volume (m^3)	Volume change (m^3)
1	01.11.1997	670.1	
2	25.06.1998	668.6	- 1.5
3	02.12.1998	666.9	- 1.7
4	20.05.1999	668.5	+ 1.6
5	09.02.2000	687.0	+ 18.5

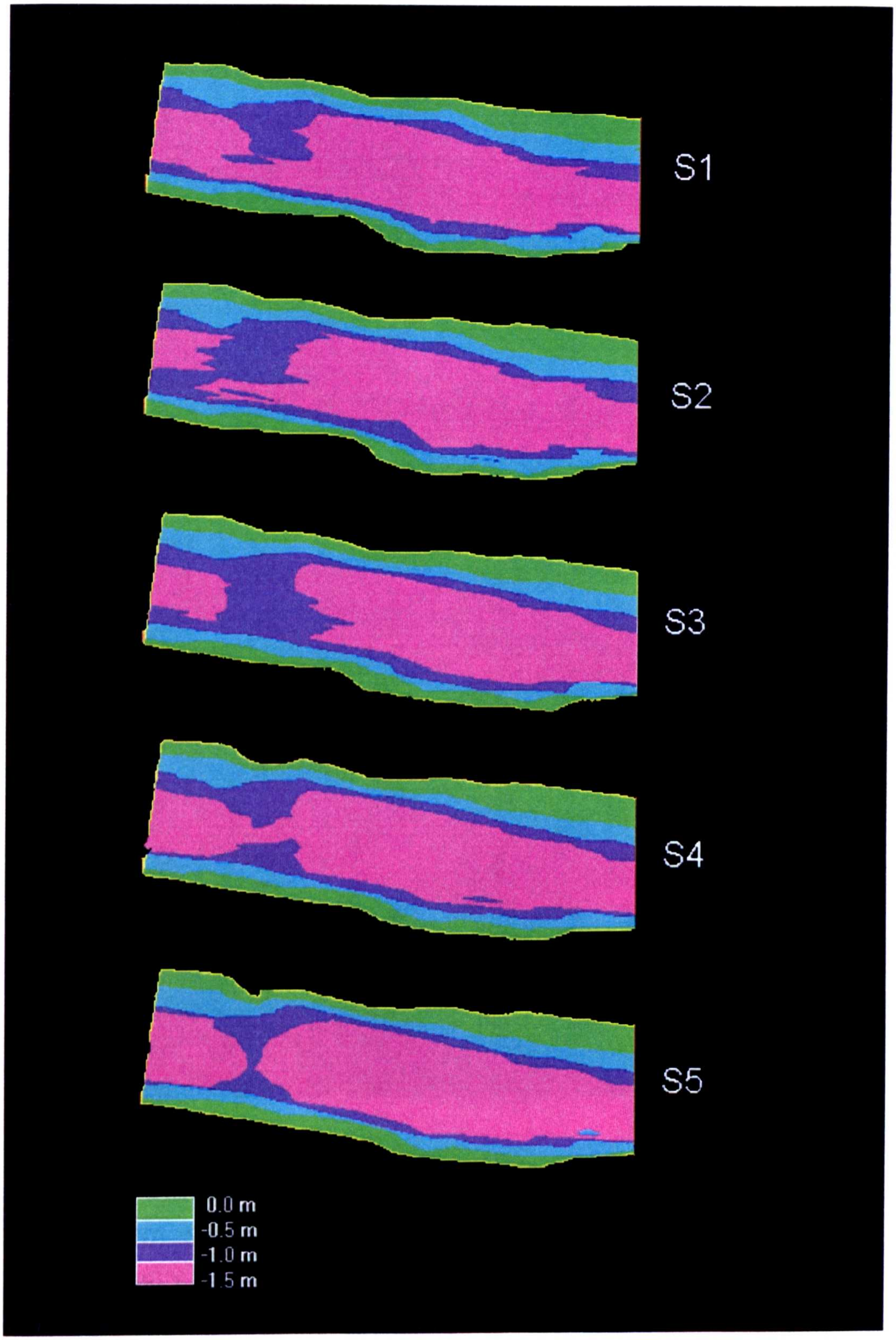


Figure 7.15 Channel DTMs of the Ddranen Ddu reach. Scale approx. 1:600. Survey numbers are indicated S1 to S5. Flow direction is left to right.

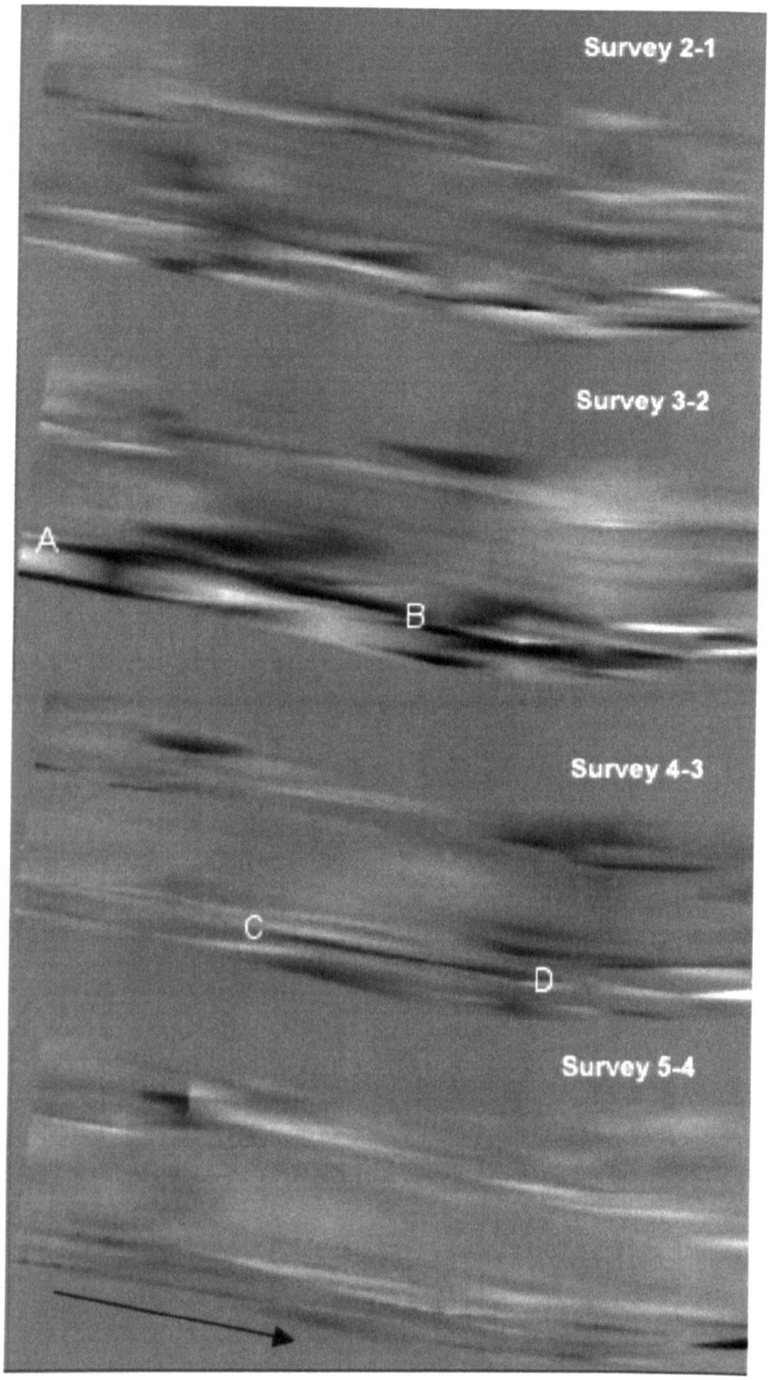


Figure 7.16 Differenced DTMs of the DD study reach for all surveys (scale approx. 1:650). Shades lighter than the background represent erosion and shades darker than the background represent deposition. Arrow indicates flow direction.

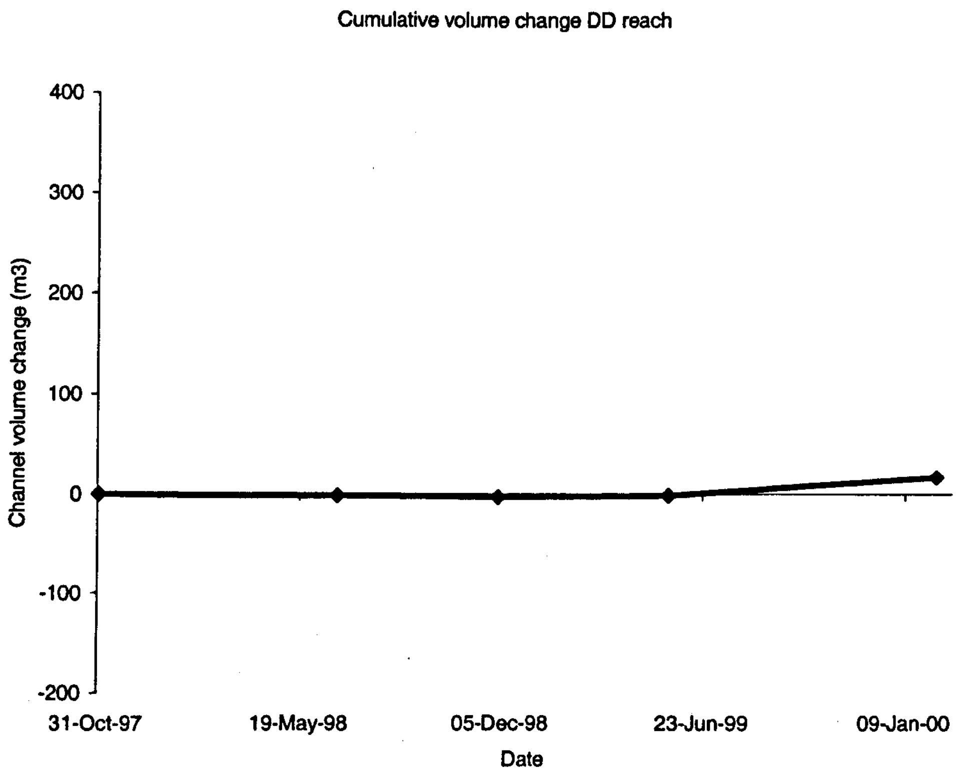


Figure 7.17 Cumulative volume change in the DD reach. Positive values show degradation.

7.4.2.4 Summary

The models presented here represent a range of qualities. The short section spacing between the sections at Ddranen Ddu (4-6 m) has produced a model of high integrity where small channel features, such as sub-meter scale bank benches, are clearly visible in the output. The low downstream variability in terms of morphology allows considerable confidence to be placed in the ability of the TIN to represent the true channel form. The reach at Ddranen Ddu shows considerable stability both in terms of bed morphology and lateral movement. Some bank slumping can be inferred from the differenced DTMs, together with some error due to survey pole tilt, however, the impact of these on the overall reach volume is small.

At Bodiach Hall the DTMs are clearly capable of allowing change in meso-scale bedforms to be analysed. The overall channel form is represented well, particularly in areas where the section density is increased. The majority of morphologic change can be traced to the development of point bars PB3 and PB4, with PB3 associated with the largest rates of lateral change. Between PB3 and PB4 the channel has maintained general morphologic stability accompanied by localised fill of isolated pools and the removal of a low-relief lateral bar. Overall the reach volume change suggests degradation.

The model sequence at Chapel View lacks the integrity seen at the other sites which has made morphologic interpretation extremely difficult. This is in direct response to the lower section density used to survey this reach and the high downstream morphologic variability of the channel. The planform maps suggest bar form deposition, particularly at PB1, however differenced DTMs are unable to wholly confirm this. The calculated volume changes within the reach are also adversely affected by the low section density and must therefore be treated with extreme caution.

7.5 Measurement and interpolation error estimates

7.5.1 Assessing the quality of the acquired interpolated surface.

In the context of an interpolated surface generated using a TIN applied to manually surveyed data points, quality may be decreased by a combination of point acquisition, gross or systematic surface errors. Point acquisition errors control the precision in the coordinate data at each point. They result from variation in the point attribute values for measurements taken under the same conditions and are detectable by repeat survey. Gross errors control the broader reliability of all of the coordinates. They are a result of blunders made by the surveyor or errors due to equipment malfunction (Torlegard, 1980). Survey density errors arise from weaknesses in the ability of the surface model to represent the true surface.

In the context of the total station survey procedure used in this study some of the sources of error are detectable whilst some are not. These error sources are summarised in table 7.5. Whilst gross error is undetectable, the use of repeat surveys has allowed attempts at point acquisition error quantification to be made. Moreover, survey density error has been investigated via comparisons of interpolated surfaces generated using a variable survey point density. The sources of random and survey density error are discussed in further detail below:

1. Point acquisition error:

This is the accuracy with which the X, Y and Z attributes of each discrete data point can be fixed. In the case of a channel volume calculated from TIN interpolated discrete data which have been arranged in section lines of fixed position (i.e. variability in the X and Y coordinates have been constrained) and where any tilt of the surveying target has been minimised by the use of a level on the target pole, the random error in the Z direction is of primary importance in respect of channel volume estimates.

Table 7.5 Sources of error in a TIN interpolated surface applied to manually surveyed data points.

Error type	Cause of error	Detection
Point acquisition: determines precision	Measurements associated with data acquisition. Controlled by: • sophistication of measuring instrument • accuracy of the placement of the survey staff	Detectable via repeat observation
	• target pole tilt	Undetectable
	• mistaken target positioning by the operator	Undetectable
Gross: determines overall reliability	• Incorrect measurement of instrument or target height during survey station setup • Inaccurate relocation of instrument position relative to fixed benchmarks for each re-survey	Undetectable in this study where repeated setting of survey equipment has not been undertaken
Survey density: determines accuracy of the representation	• Linear surface interpolation achieved by TIN between the discrete survey points	May be investigated by increasing survey point density

Where a break of slope occurs or where surface gradient is high (i.e. in a cross stream direction, particularly in channels constrained by vertical banks), only modest variability in the X and Y coordinates may produce a substantial change in the Z value. This will be introduced to the model as systematic error and requires that the relationship between gradient and elevation error should be investigated. Where the channel gradient is low small changes in the X and Y directions will have little impact on the resulting TIN and as such may be ignored. Bed load size and sorting within the reach will also have an impact on point acquisition error. In areas of large grain size, or high bed roughness the error is likely to be greatest because the pole will sometimes be placed on the top of individual pebbles and at other times it will be placed between pebbles.

2. Survey density error:
Influenced by the spatial density of data points (known as nodes in the TIN) and hence the quality of the linear surface interpolation applied between them in the construction of a TIN. In river channels, where many of the channel surfaces are curved, error will not be constant between nodes located at major breaks of slope. The greatest elevation error will occur at the midway point between the nodes and will be positive for concave surfaces and negative for convex surfaces. Towards the nodes the error will decline in a regular manner – hence this error can be thought of as systematic. The TIN can better represent surfaces if the spatial density of the nodes is increased.

7.5.1.1 Estimating point acquisition error - rationale

Volume change in gravel-bed rivers occurs as a result of erosion or deposition of material onto the river bed, where there are few major breaks of slope and surface gradient is generally low (assuming no over bank deposition). Consequently, it is the error in elevation of the portion of the interpolated surface which represents the river bed which is of primary importance in affecting channel volume. As mentioned previously, the benchmarked section surveying strategy used in this study reduces variability in the X and Y coordinates of the data points between successive surveys. As a result, the percentage error in the volume of the study reach, calculated by TIN interpolation between fixed cross-profile points, ought to be extremely similar to the percentage error attributed to measuring the bed elevation at each of the points (pers. comm. Ferguson, 2000).

In gravel-bed rivers containing coarse bed material, inaccuracy in the elevation of a point can be due to either error in the surveying equipment used, the ability of the staff operator to keep the staff or target held level, or whether the surveying staff or target is placed on top of or between individual grains on the bed. Clearly, the larger the grain size of the bed material the greater potential for error. Likewise, the potential for measurement error is intuitively greater on steeply sloping sections of bed than flat sections. To fully determine the impact of these variables on the TIN and resulting channel volume estimation, the data points require re-surveying several times during a period of no channel change (i.e. no competent flows). Subsequently, TINs for each of the re-surveys can be constructed and compared in terms of their volume. However, such an approach was logistically impossible in this project due to the fact that each study reach commonly required at least two days to gather one set of survey data. If only five survey repeats were undertaken at each of the study reaches, the fieldwork requirement would be close to five continuous weeks. In a river such as the Trannon, which displays a particularly flashy flow regime, it is difficult to envisage a window of that length where no competent flow, and hence morphological change, occurs.

In an attempt to gain a handle on the impact of detectable point acquisition error the repeated survey requirement was reduced to a single cross-profile. The cross-profile, located in the Bodiach Hall study reach and made up of 16 data points, contained zones of both steep and flat bed gradient. It was repeatedly surveyed during a period of low flow to provide a data set from which elevation error at each point could be estimated and subsequently applied to each of the cross-profile points used to construct the DTMs.

7.5.1.2 Procedure

A cross-profile was established orthogonal to the bank in the Bodiach Hall reach. A taught rope was stretched between two pegs to guide the total station target. On the first survey iteration the cross channel location of each measurement taken was marked on the rope with a piece of tape, thereby ensuring subsequent measurements were taken at approximately the same coordinates on subsequent runs. The cross-profile survey was repeated a total of 15 times. The standard error of the mean elevation at each point on the bed was computed. Points on the floodplain were ignored. The mean standard error of all these points was subsequently calculated to give a mean estimate of the bed elevation standard error for the section.

The mean standard error was used to calculated 95 % confidence limits for the volume of each of the reach TINs in turn, under the assumption that the mean standard error for the repeated cross profile was representative of all cross profiles at all study reaches. This was achieved by dividing the mean standard error by the root of n, where n is the number of survey points on the river bed at each reach survey, and multiplying the resulting point elevation standard deviation by the area of the channel. This gave the reach volume confidence interval for the survey. By multiplying the confidence interval by 1.96 the 95 % reach volume confidence limit was generated (equation 7.2).

$$CI = \left\{ \left[\left(\frac{SE}{\sqrt{n}} \right) A \right] 1.96 \right\}$$

(equation 7.2)

where

CI = 95 % confidence limit (m^3)

SE = mean standard error of the repeat survey points

n = number of points on the river bed survey

A = the channel area at each survey (m^2)

7.5.1.3 Results

The cross profile, mean bed gradient, frequency plot skewness and standard error at the surveyed points are presented in figures 7.18 and 7.19 and table 7.6. Standard errors are low with a maximum standard error of ± 0.0062 m occurring at 17.6 m and a mean standard error for the 16 points of ± 0.0035 m. Skewness varies throughout with the 16 surveyed points which contain both positive and negative values, thereby indicating that the frequency of over or underestimation of the mean elevation value is non-systematic. The relationship between standard error and bed gradient is not significant. The regression line indicates a very weak positive relationship and possesses a very high degree of residual scatter about the regression line. Consequently, it would appear that measurement error, at least at this section, is influenced predominantly by bed material and the accuracy of the total station rather than the section geometry. It also suggests that gradient-induced systematic error is minimal in the TINs.

Important channel parameters and the volume 95 % confidence limits for all surveys are presented in table 7.7. The low mean standard error value generated from the repeat cross profile, combined with the high number of survey points, has produced a low 95 % confidence limit for each of the surveys. The 95 % confidence limit, relative to reach volume, is extremely small. It is greatest for survey CV 6 where it represents only 0.04 % of the total reach volume. This reflects the accuracy of the total station survey equipment used in this study.

Volumetric changes between surveys have been calculated via simple subtraction of one volume against another where each of the two volumes have been calculated from the TIN relative to a common base plane akin to the floodplain height. No conversion of the interpolated surface from a TIN to a grid format (which can introduce further, systematic errors) was undertaken to achieve the volume change value. Consequently, propagation of error from one survey to another can be

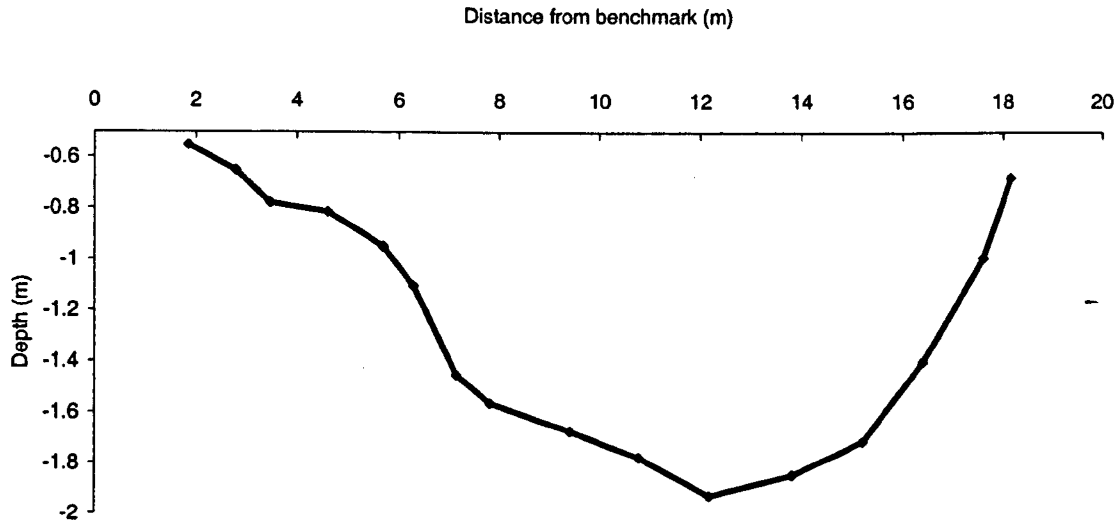


Figure 7.18 Plot of the cross profile used in the assessment of point acquisition error.

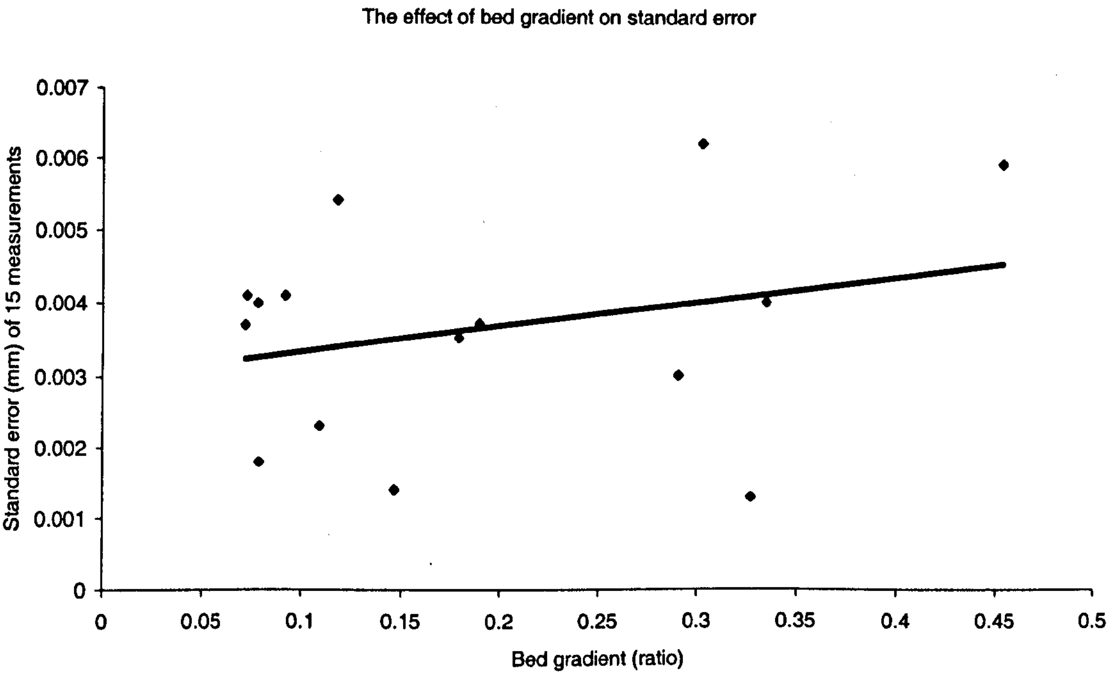


Figure 7.19 The relationship between bed cross stream bed gradient and point standard error.

Distance from benchmark (m)	Mean depth (n=15) (m)	Skewness	Mean bed gradient (ratio)	Standard error (n=15)
1.85	-0.55	0.16	0.189	0.0018
2.79	-0.65	0.05	0.327	0.0013
3.47	-0.78	0.90	0.147	0.0014
4.60	-0.81	1.78	0.109	0.0023
5.70	-0.95	-0.16	0.079	0.0018
6.30	-1.11	1.97	0.190	0.0037
7.16	-1.46	-0.67	0.334	0.0040
7.82	-1.57	-0.28	0.290	0.0030
9.40	-1.68	-0.68	0.118	0.0054
10.76	-1.78	1.07	0.073	0.0041
12.16	-1.93	0.80	0.092	0.0041
13.79	-1.85	0.07	0.078	0.0040
15.20	-1.72	0.05	0.072	0.0037
16.39	-1.41	-3.22	0.180	0.0035
17.60	-1.00	-3.68	0.302	0.0062
18.15	-0.68	0.21	0.454	0.0059

Table 7.6 Mean depth, depth skewness, bed gradient and standard error at each point surveyed on the repeat cross profile line.

Survey	Channel area (m^2)	Number of survey points	Channel Volume (m^3)	95 % confidence limit (m^3)	Quadratic sum (m^3)	Mean reach survey point density ($m^2 point^{-1}$)
BH 1	7196	1488	9893.1	± 1.32		
BH 2	7131	1307	9874.7	± 1.35	± 1.88	
BH 3	7229	1140	9849.7	± 1.47	± 1.99	
BH 4	7409	1198	9959.4	± 1.46	± 2.07	6.25
BH 5	7235	958	10045.2	± 1.60	± 2.16	
BH 6	7289	927	10231.7	± 1.64	± 2.29	
CV 1	2921	407	3474.6	± 0.99		
CV 2	3018	298	3384.2	± 1.19	± 1.54	
CV 3	3019	292	3432.7	± 1.21	± 1.69	
CV 4	3115	278	3378.0	± 1.28	± 1.76	10.54
CV 5	3059	227	3329.3	± 1.39	± 1.88	
CV 6	3040	222	3324.1	± 1.39	± 1.96	
DD 1	430	241	670.1	± 0.19		
DD 2	433	262	668.6	± 0.18	± 0.26	
DD 3	427	218	666.9	± 0.20	± 0.27	1.90
DD 4	435	240	668.5	± 0.19	± 0.27	
DD 5	428	166	687.0	± 0.23	± 0.30	

Table 7.7 Point acquisition 95 % confidence limits for the study site TIN volumes.

thought of as non-systematic. Therefore, error attached to the volume change between survey dates is the quadratic sum of the error of the two successive surveys. The inter-survey volume change and associated error is presented in table 7.8.

Inter-survey volume changes for all reaches exceed their 95 % confidence limits. At the BH reach all volume changes exceed the point acquisition confidence limits by an order of magnitude or more. The same is true at the CV reach except for inter-survey 5-6, where the volume change is less than three time the point acquisition error. In keeping with the BH reach results, point acquisition confidence limits are also exceeded by an order of magnitude or more at the DD reach.

Table 7.8 Inter-survey volume change and maximum error at the field study reaches.

Reach and inter-survey period	**Inter-survey volume change (m^3)**	**95 % confidence limit (m^3)**
BH 1-2	-18.4	± 1.88
BH 2-3	-25.0	± 1.99
BH 3-4	+109.7	± 2.07
BH 4-5	+85.8	± 2.16
BH 5-6	+186.5	± 2.29
CV 1-2	- 90.4	± 1.54
CV 2-3	+ 48.5	± 1.69
CV 3-4	-54.7	± 1.76
CV 4-5	-48.7	± 1.88
CV 5-6	-5.2	± 1.96
DD 1-2	-1.5	± 0.26
DD 2-3	-1.7	± 0.27
DD 3-4	+1.6	± 0.27
DD 4-5	+18.5	± 0.30

7.5.1.4 Discussion

The procedure for estimating volume error presented here is not fully satisfactory in that it attempts a simple translation of error estimated for a relatively simple two-dimensional condition into a complex three-dimensional condition. In other words, it assumes that the causes of the error at the repeated section are mirrored throughout the entire reach. Point acquisition error is heavily influenced by the bed material size, which can vary systematically in both cross and downstream directions over very localised reaches.

Indeed, one only need think of the downstream fining which occurs on gravel barforms to confirm this point. Where bed material size increases one might reasonably expect an increase in elevation error. However, detecting the impact of this error source would require information about the spatial distribution of bed material size within the reach and between reaches. This information is not available in this study and hence it is assumed that the bed material size at the repeated BH cross profile is representative of the distribution of bed material size across all reaches. This is probably a reasonable assumption for the CV and BH reaches which are separated by a downstream distance of only 500 m. However, the DD reach is 1.4 km downstream of the BH reach and 1.9 km downstream of the CV reach. Consequently, one might reasonably expect the bed material size to be slightly smaller at the DD reach meaning that by applying the BH 95 % confidence limit to the DD reach, the confidence limit has probably been slightly over estimated.

Point acquisition error is also influenced by the accuracy of the instrumentation used to survey the channel. This accuracy is intuitively variable with distance and one might therefore expect it to degenerate with increased distance between the instrument and target. However, the use of a total station which calculates position via the reflection of infra-red waves, the integrity of which do not degenerate with increased distance, means that the instrument has a common accuracy across a wide range of operating distances.

In the context of individual survey points, a major simplification has occurred in respect of the way each individual data point is reduced from an entity with an X, Y and Z attribute, each of which have a standard error value, to entity in which the error in the X and Y directions is considered to have little impact on the TIN and hence the overall channel volume, and as such they are ignored. This approach, albeit a simplification, is accepted in respect of the fact that each position of the target is constrained by the section line which effectively minimises imprecision in the target placement in the X and Y directions. Tilting of the survey target also has the potential to increase error in the X and Y directions and hence introduce significant volumetric error, particularly at major breaks of slope such as the interface between a vertical bank and the bed. However, target tilt has been minimised throughout this study by the use of a level on the target pole and its impact on the estimated volume of the channel, although not detectable, is hence considered minor.

7.5.2 The impact of increasing downstream section resolution

7.5.2.1 Procedure

It is clear from figure 7.2 that section resolution varies considerably throughout the study reaches. Moreover, evidence from the analysis of the CV TINs and DTMs suggests that the low section resolution in this reach has produced models of low integrity. The DD reach has the highest point density of 1 point every 1.9 m^2. This is almost five times greater than the point density at the CV reach (1 point every 10.54 m^2). In an attempt to estimate the impact that zones of lower section resolution, and hence point density, have had on the overall quantification of channel volume, three evenly spaced extra sections were installed between sections BH 6 and BH 7 during survey 6 (08.02.2000). This increased the downstream section density in that segment of channel (recorded as the distance between section end pegs on the outer bank of the meander) from 17 m to 4.25 m and the point density from 1 point 4.78 m^{-2} to 1 point 2.79 m^{-2}. The channel segment was then surface interpolated using a TIN and the volume recalculated in LISCAD for each section respectively.

7.5.2.2 Results

Results from the varied section resolution experiment are presented in table 7.9. The channel segment can be seen to decrease slightly in volume from a maximum of 360.9 m^3 with a 17 m resolution to 358.1 m^3 with a resolution of 4.25 m. This figure represents a decrease of the segment volume of close to 1 % - an order of magnitude greater than the impact of point acquisition error. Interestingly there was little volume change when section density was increased from 5.6 m to 4.25 m. It seems likely that the decrease in channel volume was due to the improved interpolation of the curved channel margins and curved bed surface of the gravel bar resulting from the increase in section density. In turn, this infers that the impact of enhancing the downstream section density in straight reaches of the study reach might also have less effect. This is especially true in light of the apparent low resolution downstream change in bed morphology of the lower Trannon at Bodiach Hall and Ddranen Ddu.

Table 7.9 Channel volume between Bodiach Hall reach sections 6 and 7, calculated using variable downstream section densities.

Distance between outer bank pegs	**Segment area (m^2)**	**Segment volume (m^3)**	**Point density (m^2 $point^{-1}$)**	**Volume change**
17 m	292	360.9	4.78	
8.5 m	292	359.8	3.36	- 0.30 %
5.6 m	292	358.2	2.58	- 0.75 %
4.25 m	292	358.1	2.19	- 0.78 %

The above results, when compared to the volumetric impact of point acquisition errors (maximum error = ± 0.04 %), demonstrate that the downstream section resolution is a more important parameter in obtaining accurate three dimensional channel information. The high point density at the DD reach would seem likely to represent the channel morphology well. At the BH reach the lower point density and section resolution suggest that the channel morphology is less accurately represented. Table 7.8 shows that even when survey density is increased to one point every 2.19 m^2 the volume of the channel segment changed by less than 1 %. Therefore, a figure of 1 % is used as an estimate of the error incurred by the lower survey resolution in the Bodaich Hall reach. This value is equivalent to approximately ± 100 m^3. This equates to a quadratic sum for each inter survey period of 1.41 %. At the CV reach the error is estimated to be greater due to the coarser survey resolution employed there. It is estimated as being twice that at the BH reach. Therefore, a value of 2 % is assigned and the quadratic sum for each inter survey period equals 2.82 %.

7.5.3 Summary

Inter-survey volume changes within each reach are presented in table 7.10 together with the combined estimates of point acquisition error and section resolution / point density error.

In the light of the above investigations of error it is apparent that volume change in the study reaches has rarely exceeded the estimated error. At CV the estimated minimum error is never exceeded. At the BH reach only inter-survey 5-6 sees volume change greater than the estimated error. Despite the volume change being greater than the error at DD the changes are small. Consequently, the volume of the Trannon channel can be thought of as showing little or no change throughout the study. However, the difference DTMs demonstrate that there has been change detectable at the sub-reach scale.

7.6 Sub-reach scale changes – investigating barform development and erosion response times

The above analysis of reach-scale planform maps and DTMs highlights the overriding importance of point bars as the major bed load storage units in the lowland channel. In the case of the BH reach, the development of point bar PB3 and, to a lesser extent PB4, are seen to have a significant impact on channel morphology. In the case of PB3 addition of new material forces the channel to migrate laterally, causing extremely high bank erosion rates. At PB4, the channel responds to deposition by deepening. In the CV reach the inferior quality of the DTMs makes the quantification of point bar development and associated erosion more uncertain. However, cross profile density is relatively high around PB1 and the DTMs do infer some deposition. In addition, the planform maps provide further evidence that there is significant downstream growth of PB1 which is accompanied by significant erosion of the opposite bank.

The small volume changes within the reaches between surveys imply that the volume of material deposited onto the point bars is quickly equalled by that removed by the associated bank erosion. Reach scale volume calculation relative to a common base plane does not isolate the volume of material entering storage in the point bar units over each inter-survey period. To quantify and characterise point bar growth, it is therefore necessary to analyse them as individual storage units, separate from the whole reach TINs. Therefore, regions of the TINs, covering only the point bars in the CV and BH reaches, have been assessed in terms of their TIN volumes, throughout the field study period. In addition, quantification of associated outer bank erosion at PB3 has also been investigated, allowing an example of the speed with which meander banks are able to erode in response to bar growth to be generated. To this end, the volume of deposition onto PB3 is accompanied by an estimate of the volume of material removed from its outer bank.

Table 7.10 Summary of TIN volume errors for all study reaches. Section resolution is estimated as capable of producing a TIN which adequately represents the channel at the DD reach. Section resolution at the BH reach is estimated to have produced TINs which represent the channel morphology to an accuracy of ± 1 % of the reach volume. At the CV reach the lower resolution is estimated to have induced a minimum of 2 % error in each TINs ability to represent the channel morphology.

Inter-survey	**Point acquisition error (m^3)**	**Survey density error (m^3)**	**Total error estimate (m^3)**	**Volume change (m^3) Values in bold show volume change > error**
BH 1-2	± 1.88	± 139.2	± 141.08	-18.4
BH 2-3	± 1.99	± 138.9	± 140.89	-25.0
BH 3-4	± 2.07	± 140.4	± 142.47	+109.7
BH 4-5	± 2.16	± 141.6	± 143.76	+85.8
BH 5-6	± 2.29	± 144.2	± 146.49	**+186.5**
CV 1-2	± 1.54	>± 95.8	>± 97.34	-90.4
CV 2-3	± 1.69	>± 97.1	>± 98.79	+48.5
CV 3-4	± 1.76	>± 95.6	>± 97.36	-54.7
CV 4-5	± 1.88	>± 94.2	>± 96.08	-48.7
CV 5-6	± 1.96	>± 94.1	>± 96.06	-5.2
DD 1-2	± 0.26	± 0.0	± 0.26	**-1.5**
DD 2-3	± 0.27	± 0.0	± 0.27	**-1.7**
DD 3-4	± 0.27	± 0.0	± 0.27	**+1.6**
DD 4-5	± 0.30	± 0.0	± 0.30	**+18.5**

7.6.1 Procedure

At the location of each point bar a block (a manually defined polygon isolating an area of interest) encompassing the barform was defined on the reach TIN for survey 1 in LISCAD. The block margins running along the banks were set by the location of the cross profile pegs and the cross profile lines, whilst the margin which fell in the channel followed the deepest point in the channel. The coordinate at each of the polygon block corners was recorded to ensure that the block could be defined in the same place on subsequent survey TINs, which could not be geographically linked on screen in LISCAD. The inability to geographically link meant that blocks defined on successive surveys differed slightly in their position due to the problems of locating exact coordinate positions on screen. Therefore, for each of the surveys, the block was defined five times and the bar volume, relative to the same base plane, was calculated for each block definition. This allowed 95 % confidence limits to be constructed for each volume change. Similarly, the erosion at the outer bank of PB3 was defined by the repeated definition of a block whose channel margins were exactly the same as those of the PB3 block. The bank margins were again defined by the cross profile peg positions.

In addition, volume errors resulting from both point acquisition and survey density issues (as discussed previously in section 7.5) were also determined for each bar volume. These errors were added to the block 95 % confidence intervals to provide an overall volume error estimate for each bar at each survey. These error bars are applied to the data subsequently.

7.6.2 Results

Bar volume for PB1 to PB4, relative to the volume at survey 1, is presented in figure 7.20. The pattern of bar growth can be seen to be similar for PB1 and PB3 (the two point bars at which bank erosion has also been identified). The volumes increase progressively through the first three surveys, before becoming stable in the later surveys. Moreover, the increases in bar volumes are consistently greater than the volume error estimates. Indeed, 148.1 $m^3 \pm 33.52$ m^3 of deposition occurs at PB1 between surveys 1 and 3, whilst deposition is greater at PB3 with 287.8 $m^3 \pm 16.23$ m^3 of deposition between survey 1 and survey 4. In contrast, the pattern at PB2 and PB4, where outer bank erosion in prevented by bank protection, is more complex. These bars can be seen to experience both deposition and removal of material, with the start and end bar volumes varying little. However, the poor TIN quality at PB2 and large errors associated with volume calculations at PB4 mean that the volume changes quantified at these bars should be treated with caution. The high error at PB 4 can be attributed to its long, narrow shape. This meant that the perimeter length of the assigned block was long compared to the other bars and hence the potential to incur error was increased.

Figure 7.21 presents the relationship between the volume of outer bank erosion at PB3 and the volume of deposition onto the bar in the previous inter-survey period. PB3 represents the most active barform in the CV and BH study reaches. The scatter plot can be seen to be highly linear, with the best fit line formula being $y = 0.9119x$. The volume of erosion from the outer bank of PB3 is almost equal to the volume of material deposited on the bar and the erosion response time is short – seemingly shorter than the inter-survey periods of this project. Indeed, the scatter plot serves as an example able to demonstrate the speed with which the channel of the lowland channel can migrate laterally following deposition. However, it is solely an observation at a single meander and is composed of only 5 observations. As such it should not be interpreted as necessarily representative of the entire channel.

7.6.3 Summary

Whilst the above procedures do not serve as a rigorous methodology for the understanding of the lowland meander development of the Trannon , having paid no

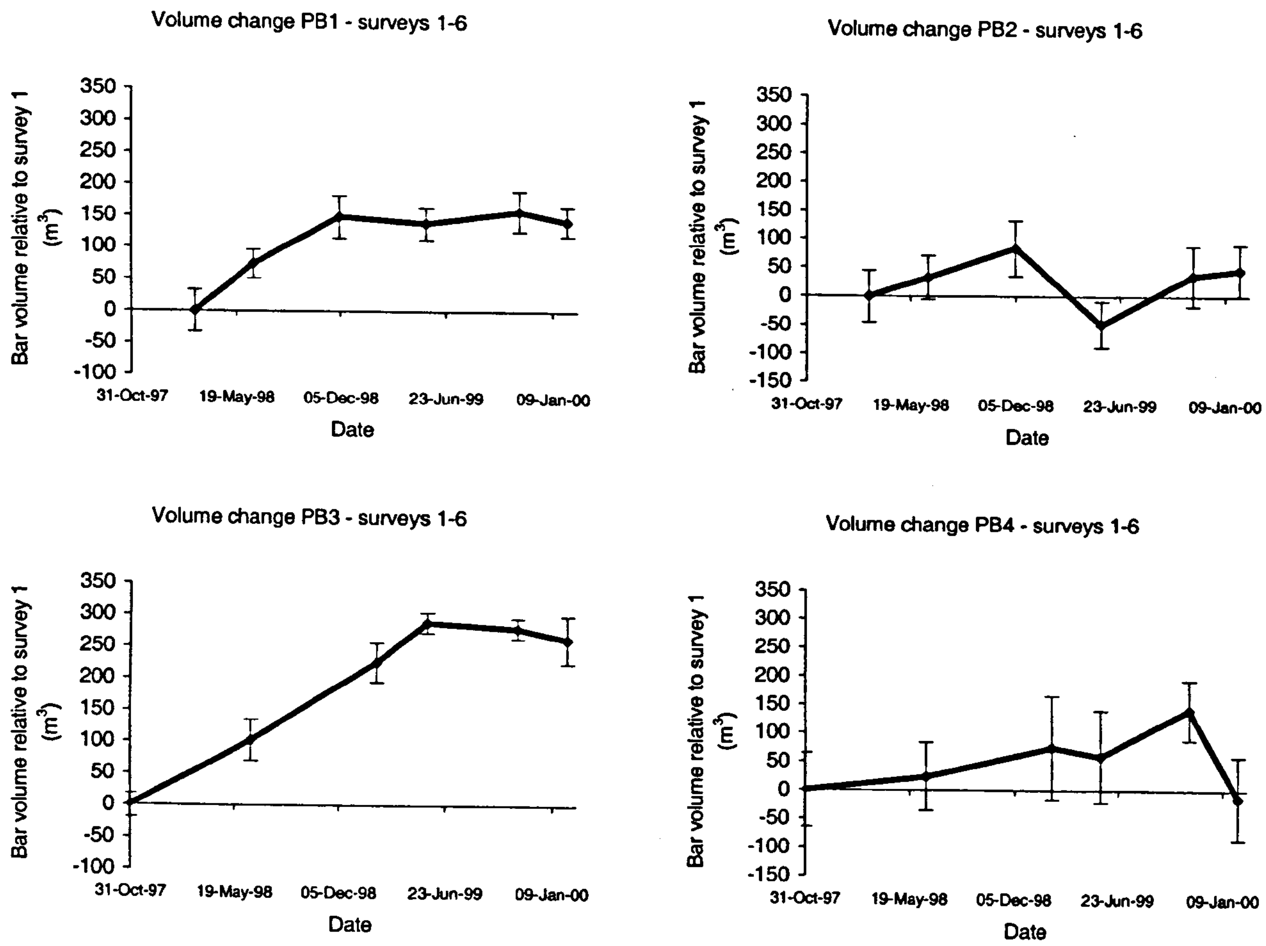


Figure 7.20 Volume change at point bars PB1 - PB4.

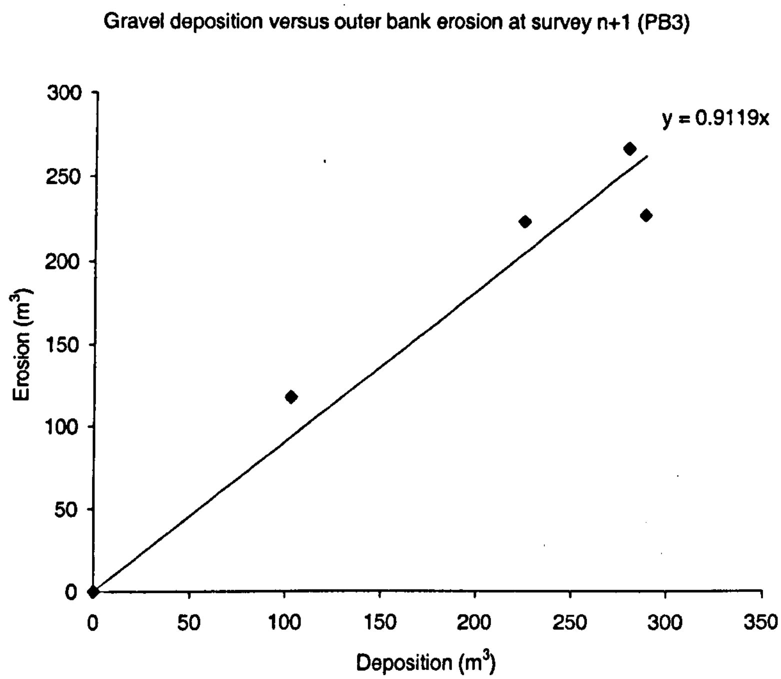


Figure 7.21 The relationship between gravel deposition of gravel onto PB3 and outer bank erosion.

attention to the specific meander geometry or fluid dynamics at each site, they do serve to show that deposition is actively occurring on point bars and that in some cases the magnitude of deposition can be very large. They suggest a difference in response between meanders which have outer bank protection present and those which do not, although high error associated with PB2 and PB4 make the extent of this difference difficult to determine. The results from PB3 serve as a single example of the apparent preference of the lowland Trannon to maintain its channel capacity in zones of deposition via lateral erosion, however it is the author's view that the process is happening in other areas of composite banks where bank protection is absent.

7.7 Spatial and temporal patterns of bed load transport rates

Studying the temporal pattern of bed load transport rates within a reach has been shown to be an excellent means of identifying sediment waves at a range of scales (c.f. Hoey, 1992; Nicholas and Sambrook Smith, 1998). Viewed from a single section spanning a channel, the passage of a sediment wave will be marked by an increase, and subsequent decrease in the bed load transport rate over a temporal series of measurements. In the case of whole reach studies in which bed load transport rates have been fully quantified throughout the reach, the passage of a single sediment wave will be expressed as a peak in bed load transport rates which progresses downstream with successive measurements. Sediment waves are of particular interest in this study due to the possibility that they may be generated as a result of forest ditch erosion in the upper Trannon catchment. Their subsequent downstream translation, and possible temporary or longer-term inclusion into bed load storage, may provide an important mechanism for lowland channel change. Consequently, bed load transport rates have been quantified for the lowland Bodiach Hall reach between October 1997 and February 2000 with a view to identifying and characterising the passage of any lowland channel sediment waves.

Several methods for quantifying bed load transport rates exist in the literature ranging from hydraulic function relationships to field techniques. Whilst hydraulic function relationships have proved useful in quantifying broad patterns of bed load transport at the basin scale, they are not easily able to cope with the spatially variable flow conditions found in rivers (Ashmore and Church, 1998) making them less useful for the investigation of transport at the reach scale or smaller. Field-based quantification of bed load transport rates have been achieved using a range of techniques including tracer particles (e.g. Carling, 1989, Wathen et al. 1997) which have produced good results. However, without the incorporation of expensive, fixed installations (e.g. Ried et al., 1980; Klingemann and Emmett, 1982; Tacconi and Billi, 1987) tracer techniques commonly require significant manual investment in their relocation and recovery between transport events.

In many recent studies (c.f. Ferguson and Ashworth, 1992; Goff and Ashmore, 1994; Martin and Church, 1995; Nicholas and Sambrook Smith, 1998) bed load transport rates have been calculated via morphological, or 'inverse' methods. Volumetric changes in sediment storage are determined, for a given time period, on the basis of change in the geometry of a cross profile survey network. The rate of change of volumetric sediment storage between successive sections may be equated with a downstream sediment transport rate increment using the following relationship:

$$\Delta Q_s / \Delta_x = -V / (T\Delta_x)$$

(equation 7.3)

where

V is the volumetric change in sediment storage between two cross profiles

Δ_x is the distance between the two cross profiles

T is the competent flow time over which the measured volume change occurs

ΔQ_s is the bed load transport rate (m^3 over time).

Whilst the above relationship determines the change in bed load transport between cross profiles, it does not provide an absolute value. In order to achieve this the bed load transport rate must be measured by some other, independent, means at the inflow or outflow of the study reach, and then propagated through it in an upstream or downstream direction. If no independent measurement of bed load transport exists an estimate of the *minimum* rate can be achieved by applying a condition of non-negative transport at all cross-profiles (Griffiths, 1979). It is the application of this basic relationship, and a non-negative transport condition, which has formed the basis by which bed load transport has been investigated in this project. A non-negative transport condition has been applied in this project therefore all figures quoted represent the minimum transport rate and do not account for material which moves through the reach without affecting the morphology.

7.7.1 Quantifying reach-scale bed load transport rates – procedure

Bed load transport rates were quantified only for the BH study reach. Quantification of bed load transport rates in the CV and DD was rejected. The reasons for this were two-fold:

1. The relatively short length of the DD reach and the relatively long inter-survey period meant that the potential for bed load to enter and leave the reach without being detected by cross-profile re-survey was high. This would mean that bed load transport rates could be seriously underestimated in this reach.
2. Whilst the length of the CV reach was considered acceptable, the relatively coarse spacing of the cross profiles meant that bed morphology at the meso-scale was not well represented by the survey. Hence, the potential of the survey to fail to detect changes of bed load storage in meso-scale bed forms (a potentially important store) was high. Consequently, whilst very large scale changes in bed load transport rates (such as those cause by macroslugs) would most likely be detected, events at a lesser scale would probably be missed.

Volume change between two neighbouring cross-profiles was achieved by isolating them as a separate file in LISCAD, and applying a TIN to interpolate the inter-profile surface. The left hand peg marking the start of each cross-profile was arbitrarily set to be at 0 m. The TIN convex hull was taken to be the cross profile lines themselves. The volume of the inter-profile area, for each inter-survey period, was calculated from the TIN, relative to a base plane set at 0 m. Subtraction of subsequent inter-survey volumes allowed V to be determined. Section spacing is non-parallel in the Bodiach Hall reach meaning that x was determined as the straight line distance between the mid point of neighbouring section lines.

Bed load transport is only able to occur during periods of competent flow. Therefore, the total period in which competent flow occurred during each inter-survey period had to be determined. This was done using the Caersws DMS record in which the entrainment threshold at Trefeglwys is represented by a DMS value in excess of 0.89 m. The number of days during each inter-survey period in which DMS > 0.89 m at Caersws was assigned as T in equation 7.2.

7.7.2 *Quantifying reach-scale bed load transport rates – results*

Volume change, stage records and the bed load transport rates for the BH reach are presented in figure 7.22 to figure 7.26. Inter-survey periods 1-2 and 2-3 (31.10.1997-19.06.1998 and 20.06.1998-15.02.1999) do not contain bed load transport rate data between 70 m and 224 m from the upstream boundary of the study reach due to the fact that cross-profile surveys were only undertaken in this locale from survey 3 (15.02.1999) onwards. The condition of non-negative bed load transport has been applied at 319 m during inter-survey period 4-5. Results have been presented in kg day rather than the conventional kg s^{-1} due to the relatively low rates of bed load transport in the lowland channel. Bed load is assumed to have a bulk density of 2650 kg m^{-3} and a porosity of 30 %.

The mean minimum magnitude of the bed load transport rates at each day above the threshold stage is 330 kg day^{-1} – equivalent to 0.0038 kg sec^{-1}. This figure was primarily set by the deposition of sediment at 320 m over the inter-survey period 4-5, in which a deep pool was almost entirely filled with bed load. The event produced a negative transport rate of –328 kg day^{-1} in the unadjusted calculations. The impact of the fill can also be seen in the differenced DTM for this period. Such low bed load transport rates strongly implies that bed load transport in the lowland Trannon is supply, rather than capacity, limited.

The existence of sediment waves is difficult to demonstrate clearly. Certainly there is no evidence to support the passage of a macro-scale wave during the study period. However, there are several zones in which the channel volume is shown to increase and subsequently decrease – indicative of temporary bed load storage followed by its subsequent re-transportation. Examples of such patterns are marked A and B on the figures 7.24 and 7.25 respectively. However, these volume changes do not translate into well defined, temporally transient peaks in the bed load transport rate. Indeed, where peaks in the bed load transport rate do occur (i.e. peak C on figure 7.24) they are not visible on subsequent plots.

To summarise, the minimum mean bed load transport rate at the Trannon is approximately 320 kg day^{-1}. Whilst plots of volume change identify zones in which the passage of meso-scale sediment waves may exist, they seldom translate into defined peaks of bed load transport rates.

Although the spatial distribution of volume change varies considerably between inter-survey periods, the range remains comparable, being between –30 m^3 and +20 m^3. However, this apparent consistency in erosion and deposition volumes is not

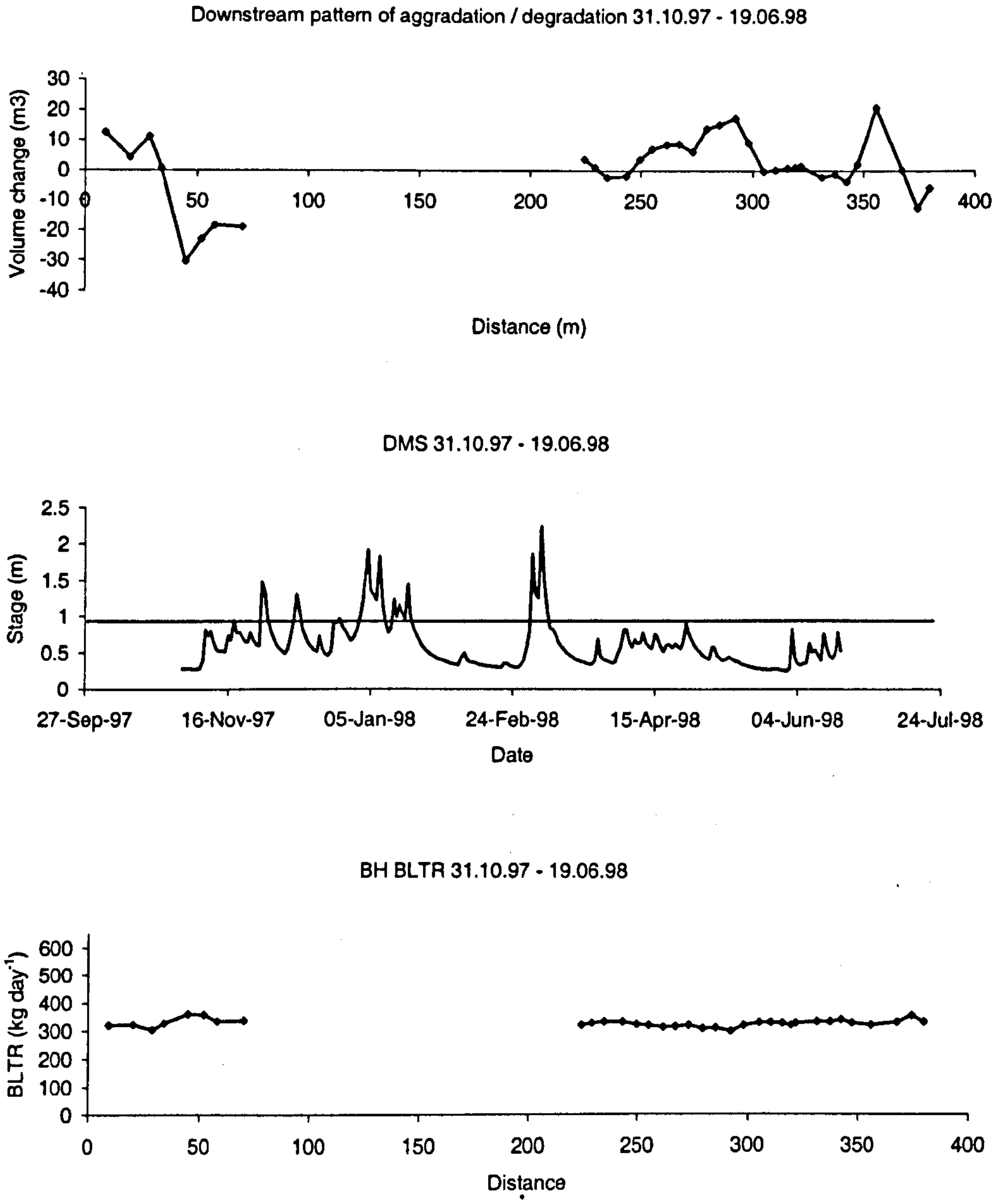


Figure 7.22 Volume change, DMS record and estimated bed load transport rates in the BH reach (survey 1-2). Entrainment threshold is shown as solid line on DMS record.

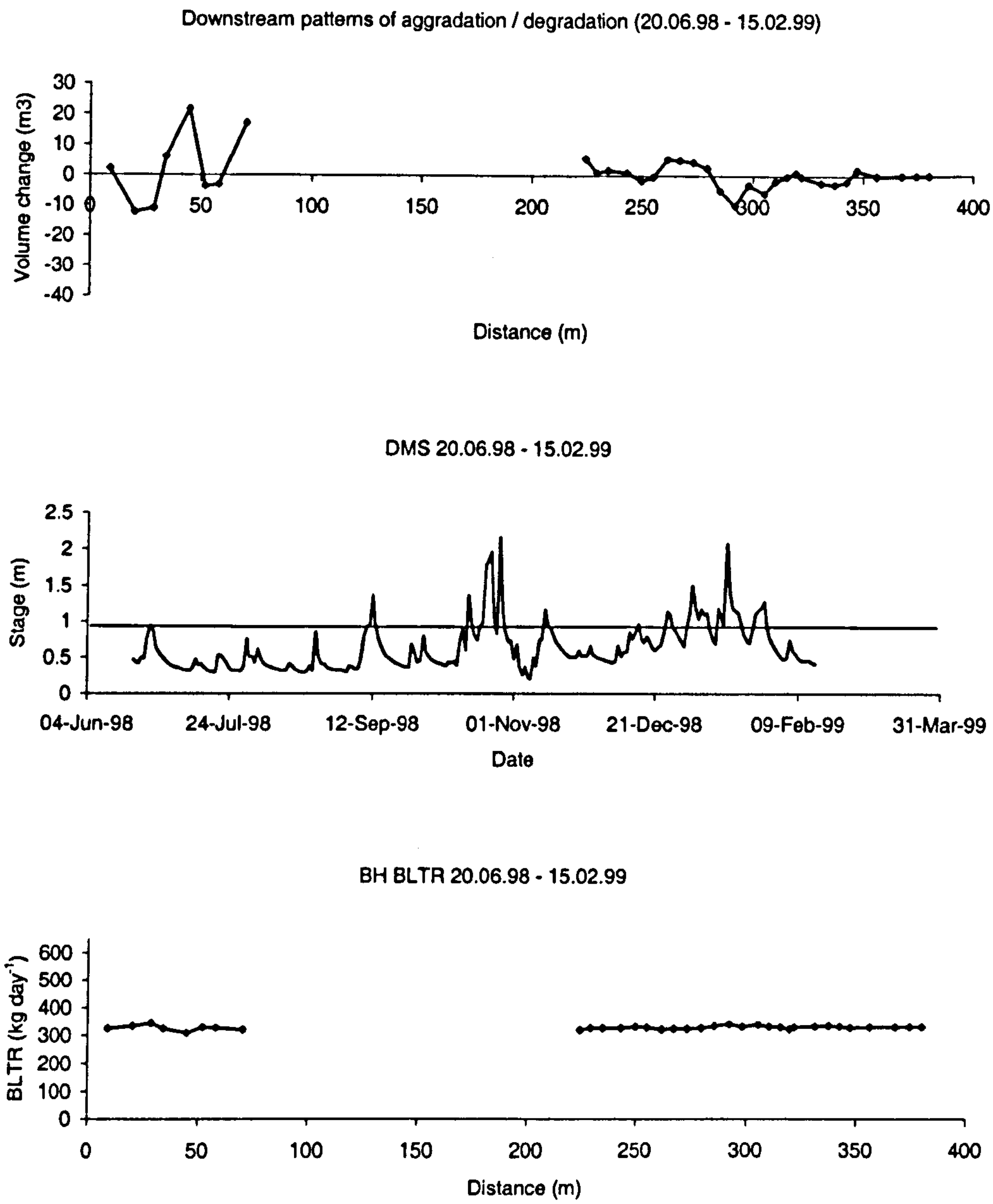
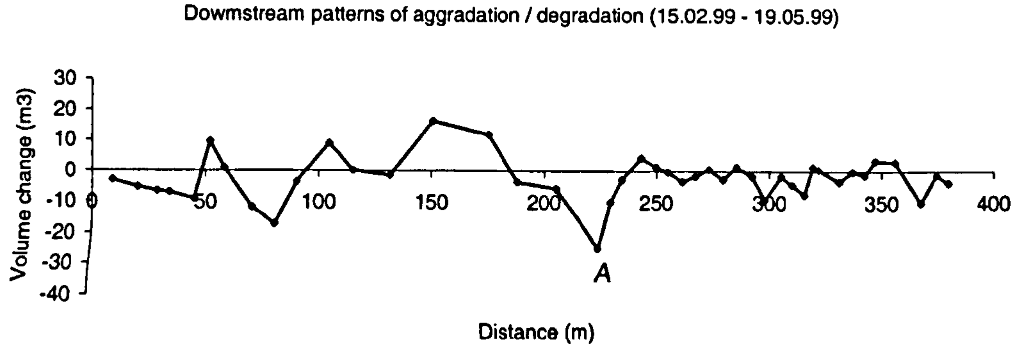
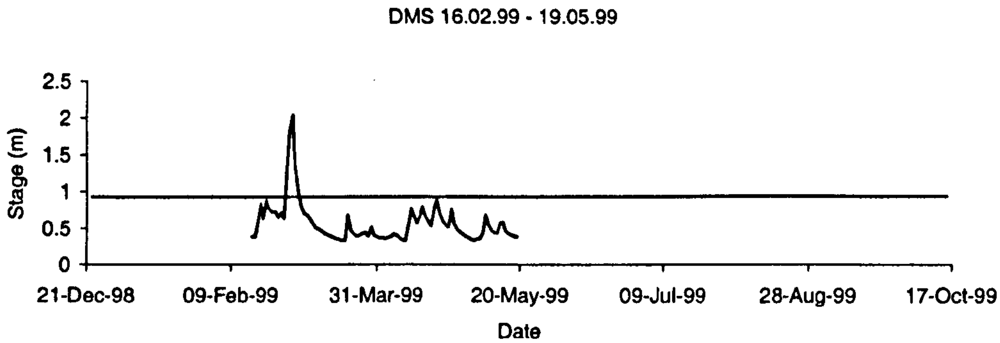


Figure 7.23 Volume change, DMS record and estimated bed load transport rates in the BH reach (survey 2-3). Entrainment threshold is shown as solid line on DMS record.





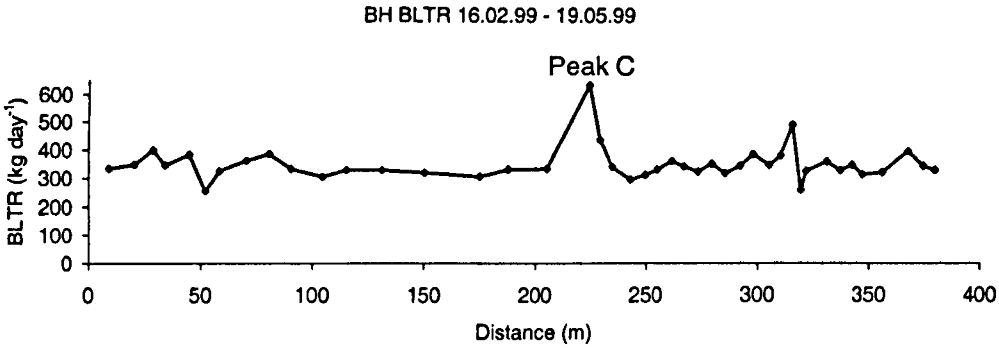


Figure 7.24 Volume change, DMS record and estimated bed load transport rates in the BH reach (survey 3-4). Entrainment threshold is shown as solid line on DMS record.

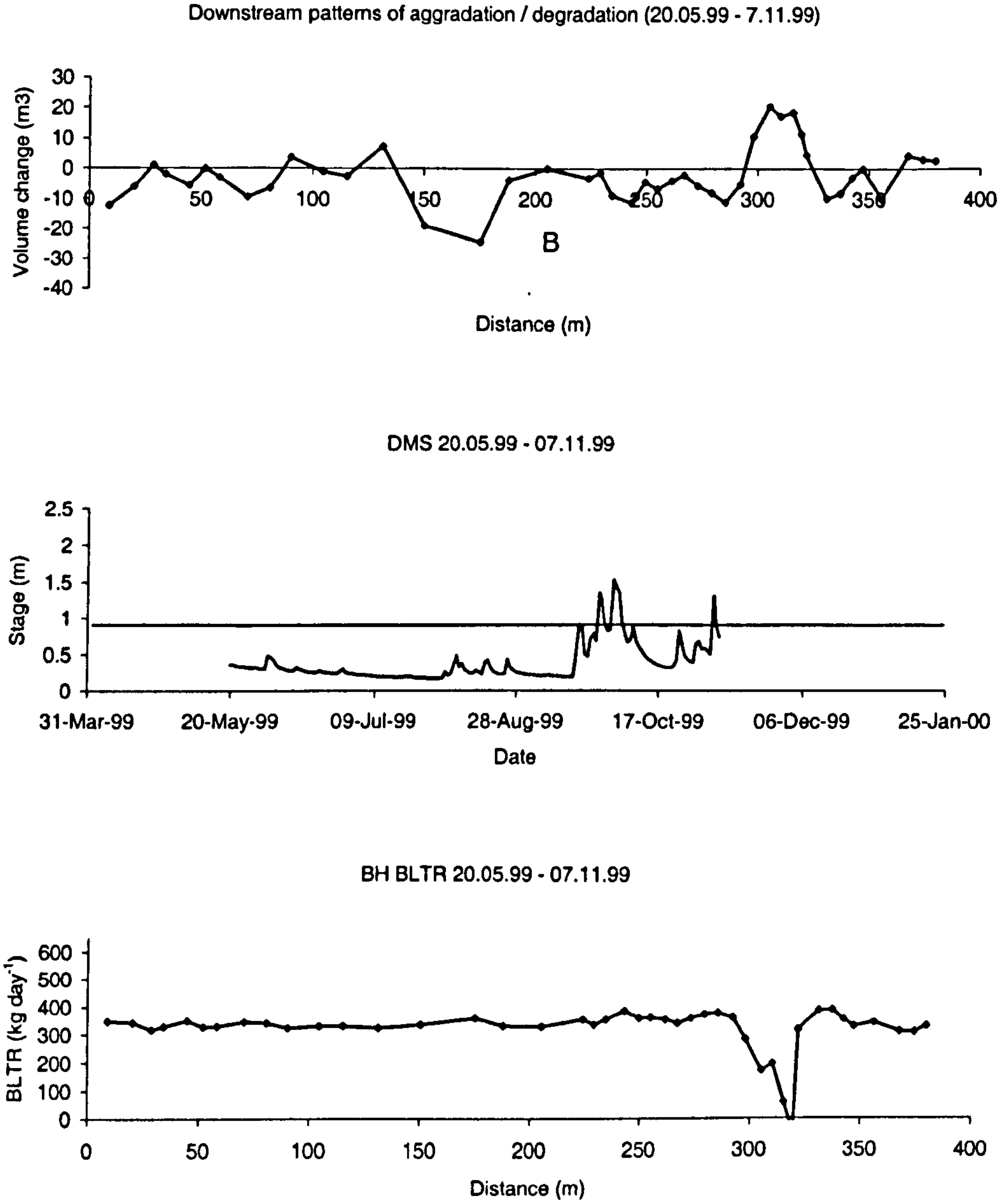
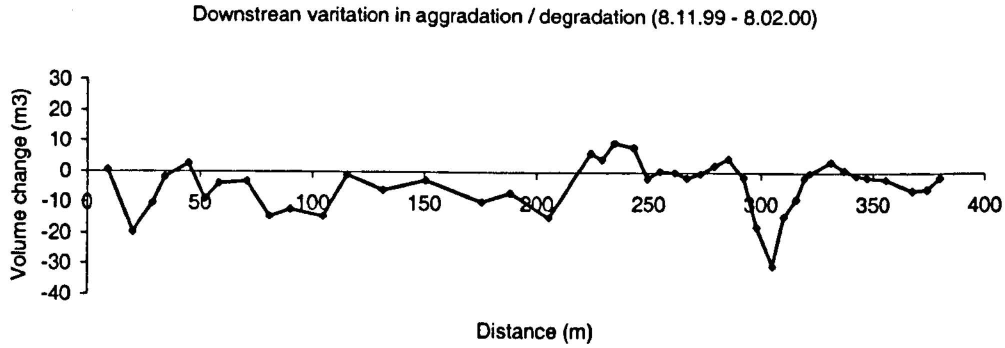
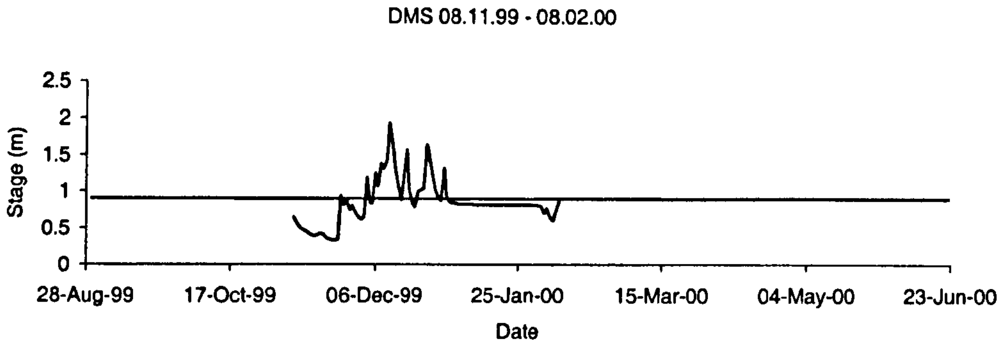


Figure 7.25 Volume change, DMS record and estimated bed load transport rates in the BH reach (survey 4-5). Entrainment threshold is shown as solid line on DMS record.





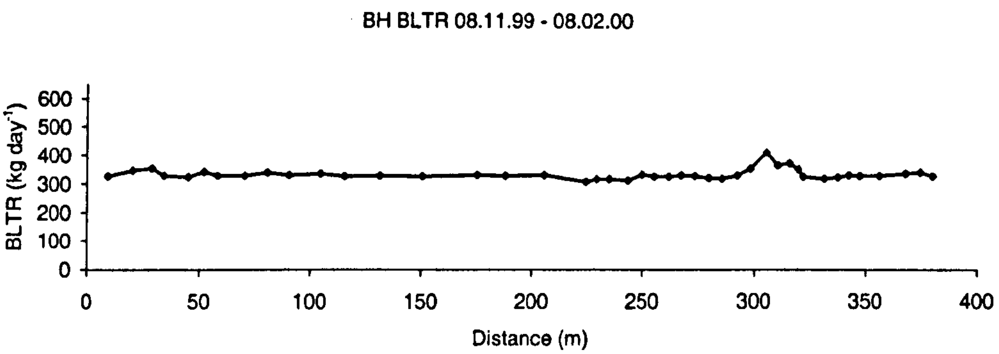


Figure 7.26 Volume change, DMS record and estimated bed load transport rates in the BH reach (survey 5-6). Entrainment threshold is shown as solid line on DMS record.

mirrored in the range of bed load transport rates. Downstream variability in bed load transport rates can be seen to be much lower in the inter-survey periods 1-2 and 2-3 than in the later periods. This is an impact of the greater frequency with which DMS exceeds the transport threshold in the longer inter-survey periods 1-2 and 2-3 which acts to reduce the magnitude of the calculated bed load transport rate. This reduction is compounded because of the high frequency with which the entrainment threshold is only just exceeded in inter-survey periods 1-2 and 2-3. Comparatively, the entrainment threshold is only exceeded by one, large magnitude hydrologic peak in inter-survey period 3-4 in which the entrainment threshold is greatly exceeded, and the resultant variability in transport rates is greater. This would imply that detailed morphologic determination of bed load transport rates would benefit from a more detailed understanding of the fractional bed load transport dynamics within the study reach.

This lack of defined, transient peaks in bed load transport rate, despite clear evidence of change in the volume of stored sediments in the channel DTMs, is perhaps not surprising given the nature of channel change presented in section 7.4. Broad investigation into planform development and more detailed investigation of the rates of gravel deposition and outer bank erosion at PB3 have showed that, in zones of unprotected composite bank material, the river will migrate laterally to maintain a channel configuration capable of containing effective discharge. This adjustment occurs over extremely short periods, possibly only the passage of a single large flood , but almost certainly over a period shorter than the temporal survey resolution. In this way the channel volume has been seen to remain largely stable despite the significant downstream transport of sediments.

7.8 Downstream variability in unit stream power

Whilst the above bed load transport analysis provides a minimum bed load transport rate estimate in the lowland channel at Trefeglwys it does not provide an assessment of bed load transport in either an upstream or downstream direction. Whilst no bed load transport rate measurements have been made in other locations in the Trannon, the bed load transport formula of Bagnold (1980) (equation 2.1) suggests that a basic qualitative assessment may be made if the downstream distribution of unit stream power is known. This is true to the extent that in regions of high stream power bed load transport rates are likely to be higher than in regions of low unit stream power for bed load of a certain size.

7.8.1 Determining unit stream power – procedure

The unit stream power formula for bankfull discharge is:

$$\Omega = (\rho g Q S)/w$$

(equation 7.4)

where

Ω = unit stream power (W m^{-1})

ρ = the density of water (1000 kg m^{-3})

g = acceleration due to gravity (9.81 m s^{2})

Q = bankfull discharge (cumecs)

S = channel gradient (m m^{-1})

w = bankfull channel width (m)

Gross stream power can be calculated:

$$\Omega_{gross} = \Omega w$$

(equation 7.5)

The density of water and gravitational acceleration are constants, therefore the variables required to compute unit stream power are discharge, slope and bankfull width. If one wishes to determine bankfull unit stream power at several locations downstream one requires knowledge of these variables at each location. Gaining channel gradient and bankfull width data is straightforward, however gaining bankfull discharge data, especially on an ungauged river such as the Trannon, is more difficult. To overcome this problem bankfull discharge has been derived using William's (1978) empirical formula which relates channel cross section area, gradient and discharge:

$$Q_b = 4.0A_b^{1.21}S^{0.28}$$

(equation 7.6)

where

Q_b = bankfull discharge (cumecs)

A_b = bankfull channel area (m^3)

S = channel gradient ($m\ m^{-1}$)

William's equation is based upon a wide ranging data set:

- 233 bankfull discharges
- discharges ranging from 0.5 to 28,320 cumecs
- cross section areas ranging from 0.7 to 8510 m^2
- slopes from 0.000041 to 0.0810 $m\ m^{-1}$

Determining the location of channel bankfull in the field is a fundamental requirement of the application of the William's formula. However, it is not a simple issue. Indeed, Williams lists 11 definitions of bankfull ranging from morphological markers (i.e. the height of the valley flat (Woodyer, 1968; Kellerhals et al., 1972) to more numerical definitions (i.e. the elevation at which the width to depth ratio becomes a minimum (Wolman, 1955; Pickup and Warner, 1976)). Outside of the incised bed rock controlled reaches of the Trannon the river has a generally well defined, relatively flat floodplain and in many locations the channel is cut into the floodplain sediments by means of vertical banks. Where these vertical banks exist the height of the valley flat is simply identified at bank top. Consequently, cross profile surveys from which bankfull discharge has been determined have been surveyed in locations where vertical banks exist.

At each location for which bankfull discharge was to be determined three cross profiles were surveyed between 10 and 20 m apart. The mean bankfull area was calculated and used as the input for the William's formula. Also at each location channel gradient was determined with a quickset level along a 100 m channel length spanning the cross profile lines. The level and staff were located as closely as possible within the thalweg, the location of which was determined visually.

7.8.2 Determining unit stream power – results

The cross profile data, calculated bankfull discharge and calculated unit stream power are presented in table 7.11.

The bankfull discharge values determined from William's formulae are difficult to independently verify in the absence of a discharge record for the Trannon. However, Leeks et al. (1988) report a bankfull flow event on the 15th March 1982 at approximately 13 km from the Trannon source which Severn Trent Water authority were able to estimate as being 25 cumecs. This figure compares well with the William's estimate at 13.2 km (23.03 cumecs) and suggests that the bankfull discharge estimates presented above are of realistic magnitude.

The data for both gross and unit stream power are presented graphically in figure 7.27. Gross stream power can be seen to peak at 8.2 km and decline quickly thereafter before stabilising at approximately 13 km. However, unit stream power has no such peak. Instead it declines at a relatively consistent rate until approximately 13 km where it stabilises. Such a pattern of consistent decline in stream power suggests a steady decline in the bed load transport rates in the upper and middle Trannon until 13 km where they stabilise.

Table 7.11 Channel gradient, bankfull area, bankfull width, bankfull discharge and bankfull unit stream power.

Distance from source (km)	S **(m m^{-1})**	A_b **(m^2)**	w **(m)**	Q_b **(cumecs)**	**Ω (W m^{-1})**
4.8	0.031	3.12	3.50		
		3.26	3.44		
		3.14	3.74		
		mean = 3.17	mean = 3.21	6.4	**547.6**
8.2	0.020	8.16	9.52		
		8.32	8.62		
		8.11	9.03		
		mean = 8.19	mean = 9.05	17.6	**382.4**
10.6	0.0115	12.56	13.45		
		10.25	11.93		
		10.95	9.34		
		mean = 11.25	mean = 11.57	22.0	**214.5**
13.2	0.0023	14.32	15.82		
		14.89	16.53		
		17.88	18.33		
		mean = 15.69	mean = 16.86	23.0	**45.5**
15.5	0.0023	18.36	17.03		
		19.22	15.35		
		19.98	15.81		
		mean = 19.18	mean = 16.06	26.2	**36.8**
18.0	0.0017	21.26	18.83		
		21.54	23.37		
		20.32	21.10		
		mean = 21.04	mean = 21.10	28.8	**22.8**

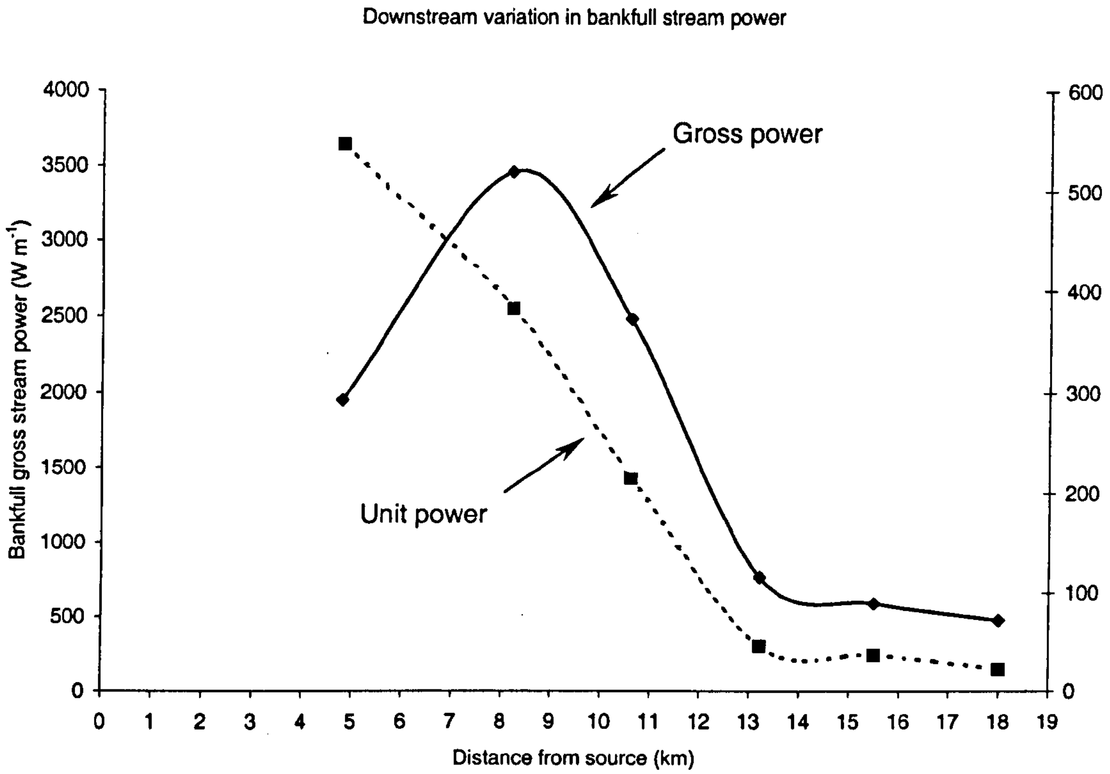


Figure 7.27 Downstream variation in unit and gross stream power.

7.9 Bed load addition from river banks

It is possible for bank erosion to be an important, if not the major, process in terms of its contribution to river sediment loads (Hill, 1973). Indeed, an early quantified study by Kirkby (1967) showed that up to 93 % of the sediment removed from the Water of Deugh in Scotland was a result of river bluff erosion. Where river banks are composed of fine grained sediments the contribution to suspended sediments is of primary importance, but where banks are composite in nature there is great potential for bed load sized material to enter the channel. In a study such as this, where sediment sources are of primary interest, it is essential to quantitatively examine the bed load inputs from river banks.

Several factors interact to influence the ease with which river banks erode (Thorne, 1982, 1990). The major ones are can be summarised as follows:

- bank height (defined as the height difference between the top of the bank and the water level at low flow)
- height of undercutting
- river bank slope
- river bank sediment type and structure
- riparian vegetation
- channel gradient
- channel mobility restricted by bedrock or channel engineering

Winterbottom and Gilvear (2000), using a GIS-based erosion prediction model based on the ideas of Graf (1984), were able to demonstrate that, of the above factors, riparian vegetation and bank composition are predominantly related to bank erosion, with bank composition being the more dominant erosion control. Of the three vegetation types assessed (grass, scrub and trees), banks upon which scrub was the main vegetation type were shown to experience erosion rates three times the mean of the other vegetation types. However, the link was thought to be indirect, resulting from the fact that scrub species are adapted to living on unconsolidated sediments typical of recently deposited floodplain units, which themselves are prone to erosion. Investigation of the relationship between bank composition and erosion demonstrated gravel and composite banks could be more than ten times more likely to experience erosion than silt banks. Composite banks were shown to be the most prone to erosion, displaying twice the erosion of wholly gravel banks.

Borehole results for the Afon Trannon floodplain are reported in Leeks et al., (1988) and presented in figure 3.2. The information shows the lowland Trannon floodplain to be a deeply scoured glacial trough, in filled with both sand and gravel. A borehole at New House Farm (SO 073914), near the confluence of the Trannon and the Severn showed an infill thickness of 68.6 m. At the Trefeglwys study reaches infill is 15.2 m deep and composed predominantly of interbedded coarse gravel and clay with gravel being the dominant component. These deep sediment fills, which form the composite banks of the Trannon, are poorly cohesive, easily eroded, and therefore offer an unusually high potential addition of bed load to the lowland sediment budget of the Trannon. Indeed, it is possible that this source of bed load could be as important a source as bed load released as a result of afforestation of the upper catchment.

In the light of the above, it was therefore necessary to map the bank material along the Trannon, between the point at which total bed rock control ceases, and the study reach begins. In this way, the relative erosion resistivity of the Trannon banks could be implied, and the extent and downstream distribution of sources of bank erosion-based bed load additions, both in and upstream of the study reach could be examined.

7.9.1 Bank material mapping – procedure

Bank material and bank characteristics were field mapped via visual assessment along 6.75 km of the channel. Bank material was mapped from immediately below the wholly bedrock controlled upper reach to directly below the BH study reach. The banks were re-evaluated whenever a significant sedimentologic or morphologic transition was seen. Both left and right banks were mapped.

The parameters recorded were chosen in the light of the studies of Thorne (1982, 1990) and Winterbottom and Gilvear (2000) and were as follows:

- Riparian vegetation
- Bank height (measured with a 10 m tape)
- Percentage gravel (where gravel clasts were defined as equal to, or larger than - 3Φ / 8 mm (small pebbles) and the percentage of gravel was determined by reference to a visual calibration card)
- Percentage sands / silts / clays (also assessed visually by reference to a visual calibration card)
- Largest visible gravel clast size (D_{max})
- River bank protection

A limitation of the above parameters is apparent in that gravel < 8 mm was deemed too small to be accurately visually assessed (a figure comparable to the minimum size which can be estimated by Wolman-type bed sampling). Therefore, the visual bank gravel estimate does not represent the total percentage bed load, which may contain significant material of < 8 mm. Instead it represents the coarse bed load additions.

Where banks were vegetated with closely spaced, established trees, they were assumed unlikely to yield significant bed load and sedimentary data were not collected. Instead, these reaches were classified as stable and vegetated. Similarly, gently sloping vegetated banks, with no exposed bank faces, were assumed to have little potential for yielding bed load and therefore sedimentary data were not collected. These banks are classified as having no exposure.

7.9.2 Bank material mapping – results

The bank material map, showing the locations of transitions in bank type, is shown in figure 7.28. The accompanying data are presented in table 7.12. Between the bedrock controlled channel and the downstream end of the BH study reach there are two regions in which active banks exist. The uppermost reach is 1.05 km in length and incorporates the Llawr-y-Glyn reach assessed using historical aerial photograph analysis. The lower reach begins 800 m upstream of Trefeglwys and extends 2.5 km to Bodiach Hall bridge. It incorporates the CV and BH field study reaches and the upstream 1.1 km of the Trefeglwys channel assessed using historical aerial photography. These reaches are separated by a 3.2 km reach, along which mature trees form the riparian vegetation and actively eroding bank faces are not apparent.

The upper active bank reach, at Llawr-y-Glyn, has grass as the dominant riparian vegetation. Bank heights are comparatively low, exceeding 1 m at only 3 sites. Sloping banks are common and occur at 5 of the 7 sites. The active bank faces can be seen to contain some gravel, however the proportion estimate rarely exceeds 10 % of the bank. Where gravel does exist, however, the D_{max} can be exceptionally large. At site 7 the largest clast recorded had a B axis of 200 mm, representing a cobble rather than a true coarse gravel. Almost no bank protection engineering has been undertaken in this reach. Indeed, approximately 100 m of bouldered banks at site 3 represents the entire protection observed. Consequently, although bed load supply to the channel as a result of bank erosion is probable in sites 1 to 7, the volume yielded is likely to be small.

At sites 8 to 26, above and at Trefeglwys, the situation is somewhat different. Riparian vegetation is more variable, with trees being a common feature. Bank heights exceed 1 m at 17 of the 18 sites and contain comparatively more gravel. The maximum percentage gravel observed was 50 %, and such high percentages were recorded at three separate locations. Banks estimated to contain 30 % gravel or more were observed at 8 sites. In further contrast to the Llawr-y-Glyn reach, bank protection was more common downstream of site 8. Protection ranged from tyre walls to tree planting, however the use of gabions and boulders were the most common bank protection measures employed. D_{max} can also seen to be smaller in the downstream active bank reach. The largest D_{max} observed was 90 mm – less than half the Llawr-y-Glyn largest D_{max}. In summary, sites 8 to 26 have a high potential to yield bed load sized material to the channel if eroded.

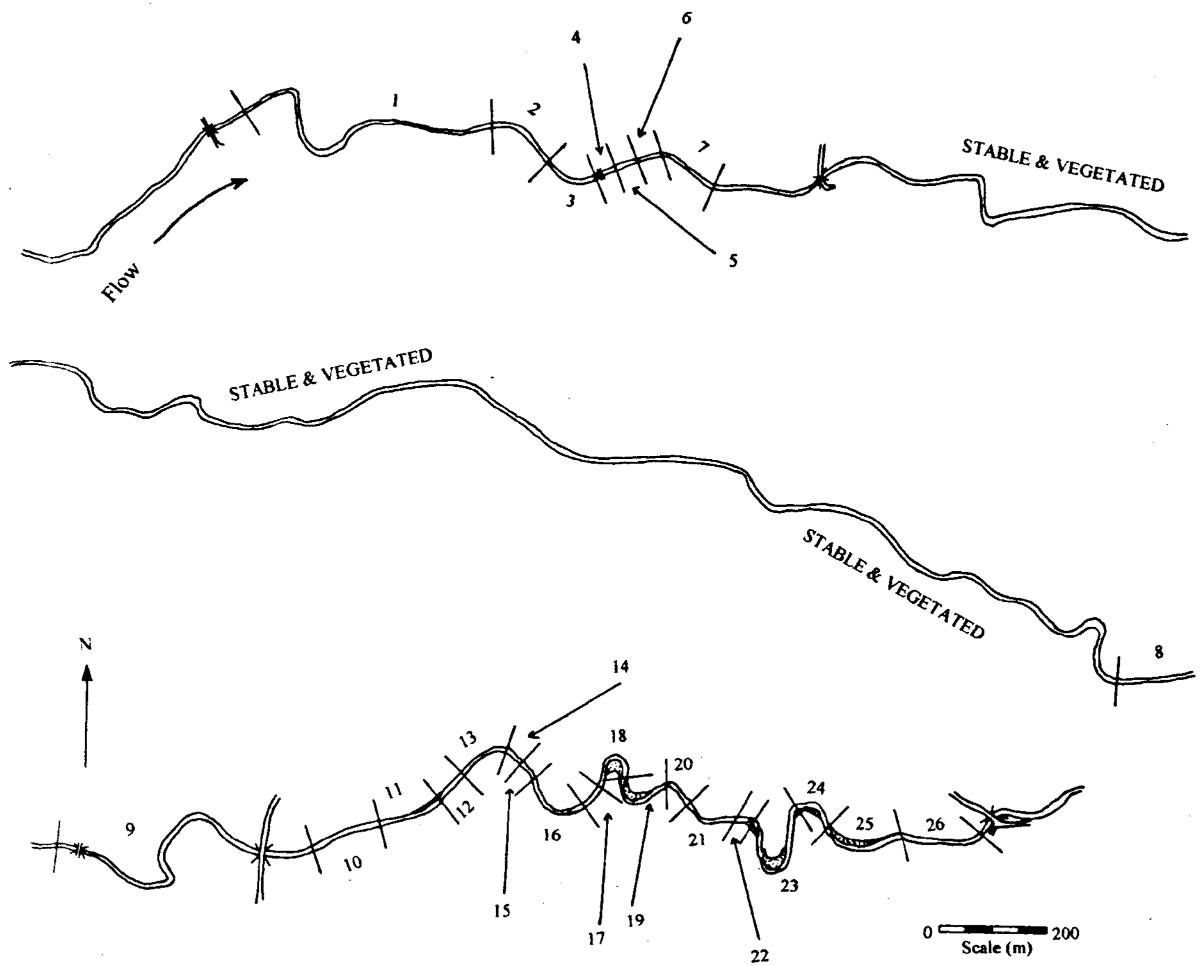


Figure 7.28 Location of bank material mapping sites. The upper boundary of site 1 is the end of full bed rock control at Llawr-y-Glyn.

Site number and bank (L=left, R=right)	Riparian Vegetation	Bank height (m)	% gravel	% fine gravel/ sand / silt / clay	D_{max} (mm)	Bank protection
1 L	Grass	0.8	5	95	95	None
1 R	Grass	Gently sloping	No exposure	No exposure	N/A	None
2 L	Grass	1.2	30	70	180	None
2 R	Closely spaced oak trees	Gently sloping	No exposure	No exposure	N/A	None
3 L	Grass	< 0.5	2	98	150	None
3 R	Grass	0.8	No exposure	No exposure	N/A	Bouldered bank
4 L	Grass	0.3	5	95	70	None
4 R	Grass	Gently sloping	No exposure	No exposure	N/A	None
5 L	Grass	0.7	20	80	110	None
5 R	Grass	Gently sloping	No exposure	No exposure	N/A	None
6 L	Trees spaced 5-10 m apart	1.5	10	90	60	None
6 R	Grass	0.6	5	95	90	None
7 L	Grass	1.2	35	65	200	None
7 R	Grass	Gently sloping	No exposure	No exposure	N/A	None
8 L	Trees spaced < 5m apart	1.5	No exposure	No exposure	N/A	Entire bank protected by tyre wall. Tyre wall was constructed in 1987
8 R	Trees spaced > 10 m apart	1.3	15	85	70	None
9 L	Trees spaced 10 m apart	1.5 – 2.5	20	80	65	None
9 R	Trees spaced 3 - 10 m	0.8 – 1.0	20	80	70	None
10 L	Grass	0.5 – 1.2	2	98	30	None
10 R	Grass	1.0	No exposure	No exposure	N/A	Extensive bouldering and gabions in place
11 L	Grass	1.0	1	99	40	None
11 R	Trees spaced 7 m apart	1.2	30	70	60	Tree planting?
12 L	Grass	0.8	20	80	70	None

Site number and bank (L=left, R=right)	Riparian Vegetation	Bank height (m)	% gravel	% fine gravel sand / silt / clay	D_{max} (mm)	Bank protection
12 R	Grass	Gently sloping	No exposure	No exposure	N/A	None
13 L	Grass	Gently sloping	No exposure	No exposure	N/A	None
13 R	Grass	1.7	No exposure	No exposure	N/A	Extensive use of gabions protects entire bank
14 L	Grass	1.0	34	66	70	None
14 R	Grass	1.3	40	60	80	None
15 L	Trees spaced < 5 m apart	2.0	0	100	N/A	None
15 R	Gravel bar	Gently sloping	No exposure	No exposure	N/A	None
16 L	Grass	1.3	25	75	75	None
16 R	Grass	Gently sloping	No exposure	No exposure	N/A	None
17 L	Grass	< 0.3	0	100	N/A	None
17 R	Grass	2.0	1	99	30	None
18 L	Grass	1.8	50	50	65	None – previously bouldered but protection has failed
18 R	Grass / gravel bar	1.5	20	80	60	None
19 L	Grass / gravel bar	Gently sloping	No exposure	No exposure	N/A	None
19 R	Low shrubs	1.6	No exposure	No exposure	N/A	Gabions protect entire bank
20 L	Grass	1.2	50	50	70	None
20 R	Grass / Gorse	Gently sloping	No exposure	No exposure	N/A	None
21 L	Grass	Gently sloping	No exposure	No exposure	N/A	None
21 R	Grass	0.8	30	70	60	None
22 L	Grass	1.5	0	100	N/A	None
22 R	Grass	Gently sloping	No exposure	No exposure	N/A	None
23 L	Trees spaced < 5m apart	Gravel bar	0	100	N/A	Extensively bouldered and trees planted

Site number and bank (L=left, R=right)	Riparian Vegetation	Bank height (m)	% gravel	% fine gravel sand / silt / clay	D_{max} (mm)	Bank protection
23 R	Grass	1.4	40	60	65	None
24 L	Grass	1.5	No exposure	No exposure	N/A	Bouldered with gabions
24 R	Grass / low shrubs	0.6	20	80	60	None
25 L	Grass	0.5	30	70	45	None
25 R	Trees spaced 10 – 15 m apart	2.0	No exposure	No exposure	N/A	Bouldered with gabions
26 L	Trees spaced <10 m apart	2.0	0	100	N/A	None
26 R	Grass	1.2	50	50	70	None

Table 7.12 Lowland Trannon bank characteristics recorded by field survey.

7.9.2.1 Bank material mapping calibration - procedure

The visually derived data presented above were calibrated via comparison with a sieved bank bulk sample removed from site number 23. The bulk sample was removed along a 25 cm wide strip extending to a depth of 25 cm and running from bank top to bank toe (150 cm). In total 0.09375 m^3 of material was removed. The sample was air dried and the gravel fractions and coarse sand (down to 1 mm sieve aperture) were sieved and weighed. The maximum clast size was also recorded. In determining the percentage of each fraction sediment porosity had to be accounted for. A porosity of 30 % was assumed for the bank material as for the bed load when converting to bulk density.

7.9.2.2. Bank material mapping calibration - results

In total 189.5 kg of material was removed giving a dry bulk density of 2.021 t m^{-3}. Assuming a mean mineral density of the bank material of 2.608 t m^{-3}, and by dividing the dry bulk density by the mean mineral density, the porosity of the bank material may be calculated. In the bulk samples analysed in this study the bank porosity is as estimated as being 22.5 %. The size distribution of the sample by mass and volume is tabulated in table 7.13. The size distribution by mass is presented in figure 7.29. The largest clast recorded had a b-axis of 73 mm.

As previously stated, the visual bank mapping assessed the percentage of gravel of 8 mm or greater. At site 23 visual mapping determined that the bank material comprised 40 % gravel of such size. The bulk sample shows that 58.6 % of the bank is gravel > 8 mm. Consequently, the visual assessment of the bank material at site 23 has underestimated the % gravel > 8 mm in the bank by 18.6 %. However, the D_{max} value recorded by both visual assessment and from the bulk sample are very similar (65 mm and 73 mm respectively). An important point raised from the bulk sampling exercise is that the visual bank assessment is limited in that it is unable to accurately distinguish the proportions of finer material in the banks which may enter the channel as bed load (i.e. < 8 mm and > 1 mm) and which the bulk sample shows may be of significant magnitude (30 %). It strongly suggests that in areas of composite banks the actual bed load addition from banks is greater than that suggested by visual bank sampling.

Sawyer (2000) reports the bed of the forested Tanllwyth, which is mainly forest ditch derived, to be composed of 21 % material between 1 mm and 8 mm based on freeze core samples to a depth of 30 mm. It would, therefore, appear that forestry also delivers significant fine bed load to upland channels. Therefore, in constructing a sediment budget for the lowland channel where bed load inputs from forestry are an input, the finer bed load fractions are also of importance.

Table 7.13 Site 23 (Bodiach Hall) bank material size distribution

Sieve aperture (mm)	Mass (kg)	% by mass
63	5.21	2.75
45	10.10	5.33
31.5	23.78	12.54
22.4	25.20	13.30
16	19.41	10.24
8	27.51	14.52
4	17.70	9.34
2	12.98	6.85
1	24.88	13.13
< 1	22.73	12.00

7.9.3 Estimating medium-term bed load yield from active gravel banks

The contemporary bank map data, combined with lateral channel movement data obtained from historical aerial photography, provide a data source from which an

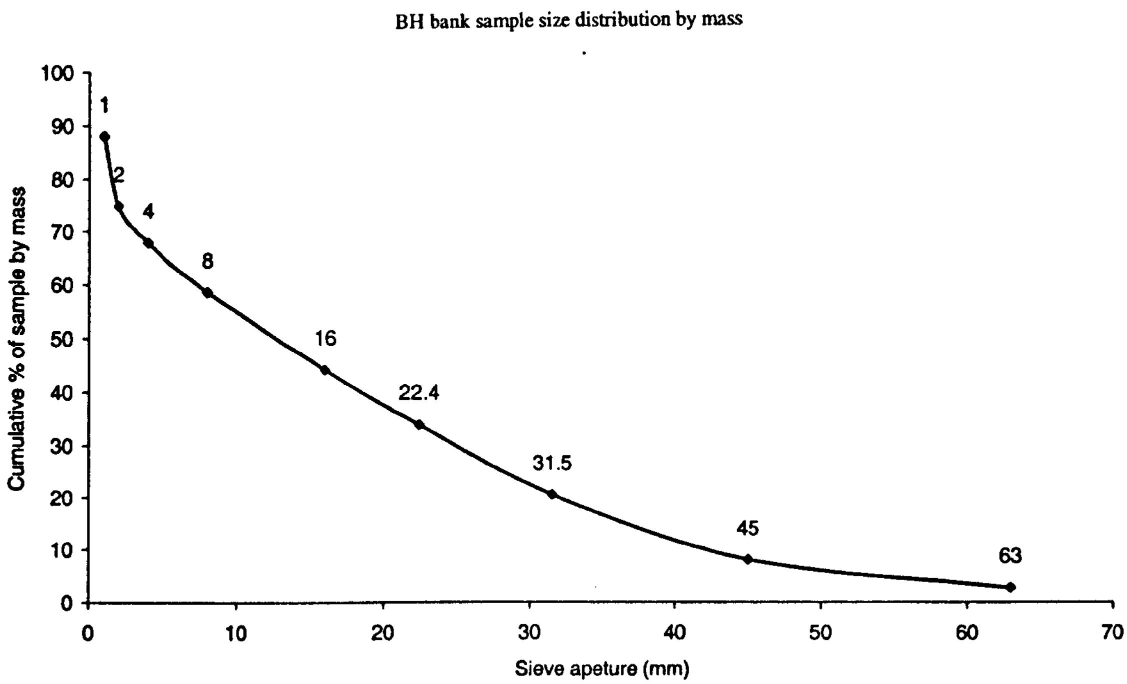


Figure 7.29 Size distribution of bank material bulk sample between 63 mm and 1 mm.

estimate of the bed load derived from bank erosion over the last 50 years can be made. Additionally, this allows a quantitative comparison to be made between the relative importance of upland catchment bed load yields and lowland reach bed load yields from bank erosion. To this end, site length and bank height data have been used to gain an estimated active bank area value for the channel between sites 1 to 7 and sites 8 to 26. Subsequently, the % gravel data (of 8 mm size or greater) have been included to allow an overall % gravel > 8 mm estimate for each of the active reaches. These data are presented in table 7.14. Left and right bank heights have been averaged at each site, as have the gravel percentages. The percentage gravel figure has been converted to a dimensionless ratio in which 0 % is equal to 0.0 and 100 % is equal to 1.0. Because bank face area and gravel area have been calculated individually for each of the sites and then combined, the longer reaches are more heavily weighted in regards to their effect on the final percentage gravel value. In this way the percentage gravel estimate provides an overall value for the active reach, assuming a uniform distribution of gravel throughout the banks. It is clear from the table that the upstream most active reach at Llawr-y-Glyn has a far lower overall gravel proportion in the banks (6.7 % > 8 mm), compared to the active reach at Trefeglwys (16.1 % > 8mm). It should also be noted that based on the above calibration, the historical coarse gravel bank yield estimates presented below are minimum values and, based on evidence from the bulk sample calibration, may underestimate the coarse gravel inputs by close to 20 %.

Table 7.15 presents estimates for historical bed load yield at the Llawr-y-Glyn and Trefeglwys active bank reaches. The estimate has been generated using the mean lateral channel movement values for the Llawr-y-Glyn reach and the Trefeglwys reach (between 13 and 14 km), obtained from historical aerial photo analysis (presented and discussed in chapter 6). By multiplying together the active bank face area (m^2) estimated from contemporary field mapping, and the total mean lateral movement (m) estimated from aerial photograph analysis, the volume of eroded bank entering the channel can be estimated for the given aerial photo interval. Application of the percentage gravel > 8 mm estimate for each reach provides a final estimate of the volume of coarse bed load entering the channel from each reach over the period of the aerial photography. This volume can then be converted to a bed load mass and used in direct comparison to the estimated upper catchment bed load yields presented in chapter 4 with Sawyers 21 % (a figure also assumed representative of the bed load between 1 mm and 8 mm yielded from the upper Trannon) subtracted by means of a simple correction for the proportion of material of < 8 mm in the bed.

The results shown in table 7.15 demonstrate medium-term bed load yield estimates which vary by up to two orders of magnitude between the Llawr-y-Glyn and Trefeglwys reaches. At Llawr-y-Glyn, low bank gravel percentages, relatively low bank heights, a relatively short active reach length and a historically stable channel mean that little gravel is estimated to have been yielded from the reach between 1948 and 1995. In total, 84 m^3 (155 t) of coarse bed load sized material is estimated to have entered the Trannon as a result of bank erosion – equivalent to just 1.8 m^3 yr^1 (3.3 t yr^{-1}). Measurement error in the aerial photography analysis means that the bed material yield could be between 0 m^3 (0 t) and 303 m^3 (561 t). However, even this maximum possible yield represents a very limited bed load source compared to the upper catchment yield of 1060 m^3 (1967 t) of material > 8 mm over the same period.

In contrast, the Trefeglwys reach is longer, has higher bank heights, a greater proportion of gravel in the banks and has experienced consistently higher rates of lateral movement than Llawr-y-Glyn, particularly since 1988. These factors have combined to produce a highly significant bank erosion based bed load yield estimate. Between 1948 and 1976 the estimated yield is 1709 m^3 (3170 t) – equivalent to 61 m^3 yr^{-1} (113 t yr^{-1}). This volume is similar to the total estimated upper catchment yield of gravel > 8 mm for the period 1949 – 2000, and far exceeds the 711 m^3 (1318.9 t) of gravel > 8 mm estimated to have been yielded from the upper catchment over the same 28 year period. However, the large aerial photograph measurement error inherent in the analysis of the poor quality early aerial photography is expressed in a large range of possible yields, from a minimum possible yield of 256 m^3 (equal to 9.1 m^3 yr^{-1} or 16.9 t yr^{-1}) to a maximum of 3162 m^3 (equal to 113 m^3 yr^{-1} or 209 t yr^{-1}).

In the period 1988 to 1995 analysis of aerial photography suggests that the lateral change rate at Trefeglwys was similar to that in the previous period (particularly between 1963 and 1976). The improved quality of the photography, however, means a lower measurement error in proportional to overall channel movement. Mean total lateral channel movement was 3.19 m between 1988 and 1995, meaning that more than 8750 m^3 of bank erosion is estimated. In terms of coarse gravel yield, 1409 m^3 (2613.5 t) is estimated to have entered the channel as a result of bank erosion. This is equivalent to 201 m^3 yr^{-1} (373 t yr^{-1}), and is more than double the estimated annual bed load yield of gravel > 8 mm from the upper catchment for the same period, although the high error means that the input range is between 6.5 m^3 yr^1 (12 t) and 94 m^3 yr^{-1}(174 t yr^{-1}).

Site number	Length (m)	Mean bank height (m)	Bank face area (m^2)	Mean proportion of gravel (ratio)	Total bank face area composed of gravel (m^2)
Llawr-y-Glyn					
1	570	0.4	228	0.025	5.7
2	130	0.6	78	0.15	11.7
3	115	0.65	74.75	0.01	0.75
4	40	0.15	6	0.025	0.15
5	45	0.35	15.75	0.1	1.6
6	40	1.05	42	0.075	3.15
7	110	0.6	66	0.175	11.55
TOTAL	**1050**		**510.5**		**34.55**
GRAVEL = 6.7%					
Trefeglwys					
8	200	1.4	280	0.075	21
9	620	1.45	899	0.2	179.8
10	130	0.93	120.2	0.01	1.2
11	110	1.1	121	0.155	18.75
12	70	0.4	28	0.1	2.8
13	90	0.85	76.5	0.0	0.0
14	30	1.15	34.5	0.37	12.8
15	40	1.0	40	0.0	0.0
16	135	0.65	87.8	0.125	10.97
17	40	1.15	46	0.005	0.23
18	100	1.65	165	0.35	57.75
19	135	0.8	108	0.0	0.0
20	55	0.6	33	0.25	8.25
21	100	0.4	40	0.15	6.0
22	35	0.75	26.3	0.0	0.0
23	275	0.7	192.5	0.2	38.5
24	95	1.05	99.8	0.1	10.0
25	110	1.25	137.5	0.15	20.6
26	130	1.6	208	0.25	52.0
TOTAL	**2500**		**2743**		**440.65**
GRAVEL = 16.1%					

Table 7.14 Estimate of the percentage gravel in the banks of the Llaw-y-Glyn and Trefeglwys reaches.

Study reach / date		
Llawr-y-Glyn: 1948-1976		
	Bank area (m^2)	510.5
	Mean lateral channel movement (m)	1.12 ± 3.29
	Mean lateral movement error range (m)	0 – 4.41
	Mean total bank erosion (m^3)	571.76
	Minimum possible bank erosion (m^3)	0
	Maximum possible bank erosion (m^3)	2521.46
	Total volume of gravel > 8 mm yielded (m^3)	**38.3**
	Minimum possible gravel > 8 mm yield (m^3)	0
	Maximum possible gravel > 8 mm yield (m^3)	168.9
Llawr-y-Glyn: 1976-1995		
	Bank area (m^2)	510.5
	Mean lateral channel movement (m)	1.33 ± 2.58
	Mean lateral movement error range (m)	0 – 3.91
	Mean total bank erosion (m^3)	679.0
	Minimum possible bank erosion (m^3)	0
	Maximum possible bank erosion (m^3)	1996.0
	Volume of gravel > 8 mm yielded (m^3)	**45.5**
	Minimum possible gravel > 8 mm yield (m^3)	0
	Maximum possible gravel > 8 mm yield (m^3)	133.7
Trefeglwys: 1948-1976		
	Bank area (m^2)	2743
	Mean lateral channel movement (m)	3.87 ± 3.29
	Mean lateral movement error range (m)	0.58 – 7.16
	Mean total bank erosion (m^3)	10615.4
	Minimum possible bank erosion (m^3)	1590.9
	Maximum possible bank erosion (m^3)	19639.9
	Volume of gravel > 8 mm yielded (m^3)	**1709.0**
	Minimum possible gravel > 8 mm yield (m^3)	256.13
	Maximum possible gravel > 8 mm yield (m^3)	3162.0
Trefeglwys: 1988-1995		
	Bank area (m^2)	2743
	Mean lateral channel movement (m)	3.19 ± 2.78
	Mean lateral movement error range (m)	0.41- 5.97
	Mean total bank erosion (m^3)	8750.2
	Minimum possible bank erosion (m^3)	1124.63
	Maximum possible bank erosion (m^3)	16375.7
	Volume of gravel > 8 mm yielded (m^3)	**1408.9**
	Minimum possible gravel > 8 mm yield (m^3)	181.1
	Maximum possible gravel > 8 mm yield (m^3)	2636.5

Table 7.15 Estimated bed load addition from Llawr-y-Glyn and Trefeglwys bank erosion 1948-2000 for gravel > 8 mm.

7.10 Chapter summary

In section 7.1 four questions were presented. The answers to these questions are presented below.

1. Is significant bed load deposition and storage presently occurring in the destabilised Trefeglwys reaches? Is this situation different in the more stable Ddranen Ddu reach?

Whilst analysis of reach volume changes suggests that little within reach bed load deposition is occurring within the study reaches, evidence from the planform maps and differenced DTMs at the BH and CV reaches strongly suggests otherwise. They show increase in both the surface area and volume of the reach point bars – a clear indication of bed load deposition. When analysed separately increase in the volume of gravel bars is confirmed, with the greatest deposition occurring around meander bends with little or no channel engineering. Some scour and fill of the bed in the BH reach can be seen, particularly in the pools, however this deposition is minor when compared to that on the point bars.

At Ddranen Ddu channel volume change is also very low, however in contrast to the BH and CV reaches there is little bed load storage evident. There are no gravel bar forms in the reach and the majority of reach volume change is due to minor changes in bed elevation. Overall, the reach shows slight degradation indicating that bed erosion rather than deposition has occurred.

2. What are the locations and magnitudes of bank erosion within the reaches? Do these patterns correspond with locations and magnitudes of bed load deposition?

From the above it is clear that bank erosion must be an extremely important process in terms of the lowland within-channel sediment budget and maintenance of the channel capacity of the lowland Trannon. This is particularly true for the BH and CV reaches. Differenced DTMs, particularly of the BH reach, show that the majority of bank erosion occurs in the region of, and in response to, deposition of bed load onto point bars. However it is also apparent that river engineering has restricted bank erosion to the channel around point bars where no outer bank protection exists. Evidence from PB3 suggests that the magnitude of bank erosion can equal the deposition of gravel onto barforms and that this can be achieved within a single inter-survey period.

At the Ddranen Ddu reach there is little bank erosion. Differenced DTMs infer some rotational slumping of the banks but there is none of the major planform change and channel migration seen in the BH and CV reaches.

3. What is the magnitude of bed load transport within the reaches? Can the passage of any bed load waves be identified in the study period?

The mean rate of bed load transport in the BH reach throughout this study was 320 kg m^{-1} day^{-1}. This figure represents a minimum value according to Griffiths (1979). Commonly bed load transport rates varied little downstream despite some locally large channel volume changes. This was due to the high frequency with which the DMS entrainment threshold was commonly exceeded between surveys. Where bed load transport rates did vary, suggesting a possible bed load wave, the peak in bed load transport was not detected downstream in subsequent surveys. Consequently, no sediment waves were detected in the BH reach during the field study period.

4. What is the contribution to the bed load of the lowland channel from bank erosion? Is it more / less important a source than that from the upland catchment?

The estimated contribution of bed load > 8mm to the lowland Trannon from combined bank erosion at Llawr-y-Glyn and Trefeglwys between 1948 and 1995 is

3202 m^3. This is more than three times the 1060 m^3 of bed load > 8 mm estimated to have been yielded from the upper catchment during the same period. It strongly suggests that lowland bank erosion of gravel and composite banks has the potential to deliver large quantities of coarse bed load to the lowland channel. This is particularly true when one considers that bank material between 1 mm and 8 mm, which is a potential source of the finest bed load fractions and was not assessed by visual bank mapping, has been shown by the bulk sample analysis to equal approximately 30 % of the bank by volume at bank mapping site 23. Indeed, bed load yielded from bank erosion is a more important source of bed load into the lowland channel sediment budget than that from the upper catchment.

8. DISCUSSION

Chapter overview

This chapter seeks to integrate the results presented in chapters 4 to 7 in the context of examining evidence for and against a causal link between the lowland instability of the Trannon and the afforestation of the upland catchment. Here the historical and contemporary channel changes of the Trannon are discussed in comparison with upland catchment bed load yields, the medium-term hydrologic setting, possible cumulative effects within the sediment routed through the channel network and lowland bed load inputs from river banks. The relative importance of bed load sources are considered and a conceptual model which explains the lowland patterns of channel instability is presented. Finally, the findings of this single catchment study are contemplated in terms of their inferences for the management of other rivers in the UK.

8.1 Upland forestry, elevated upland catchment bed load yields and lowland destabilisation – evidence for a direct causal link

As presented in section 2.13, Leeks et al. (1988) offer limited circumstantial evidence which suggests a link between lowland channel destabilisation and upland catchment afforestation and state that the afforestation of the upper Trannon may have implications for lowland channel stability. This evidence centers about the timing of increased channel destabilisation in the lowland channel. Channel instability is reported to have increased following upland catchment afforestation. Accompanying this instability are increases in gravel barform volume which suggests a possible increase in coarse sediment supply to the lowland reach. These changes occurred in the absence of any other major land use changes identified in the catchment. In the present study, the patterns of medium-term lowland channel changes have been analysed in greater detail in order to generate a more comprehensive data set from which the temporal and spatial nature of lowland channel changes can be compared to the timing of upland catchment afforestation.

8.1.1 Evidence for the implication of upland forestry

If upland forestry is to be implicated in respect of the historical pattern and timing of channel changes which have occurred in the lowland Trannon several key features should be apparent in the results from the preceding chapters. These features are discussed in the following listed sections:

1. (Discussed in section 8.1.2 and 8.1.3)

The timing of augmentation of the historical rates of lowland channel change should be lagged behind the afforestation of the upper catchment. The length of the lag time (L) should be equal to the mean bed load transport rate $Q_{b(mean)}$ in the channel between the forest and the reach experiencing the change multiplied by the distance between the forest and affected reach (x):

$$L = Q_{b(mean)}x$$

(equation 8.1)

2. (Discussed in section 8.1.4)

Evidence that historical bed load storage has increased in the reach experiencing the change should exist. The timing of the increase in storage should match the timing of the augmentation of channel change rates.

3. (Discussed in chapter 8.1.5)

A mechanism whereby non-point source forest bed load may experience accumulation in the affected reach should be present.

4. (Discussed in section 8.1.6)

The role of changes to the magnitude and frequency of discharge should be such that channel change can not be explained solely by an increase or decrease in the magnitude and frequency of competent flow events unless these are related to forest practices.

5. (Discussed in section 8.1.7)

The quantity of bed load yielded from the upland catchment should approximately equal that entering storage in the lowland reaches.

8.1.2 Lowland response lag time

Results from the analysis of archive aerial photography for the upper 1000 m of the Trefeglwys study reach at Trefeglwys (presented in section 6.9) are compared with historical channel change data obtained from historical maps and presented in Leeks et al. (1988). The data are summarised in table 8.1 below.

It should be noted that the value for mean lateral movement is generated by combining the positional change in both left and right banks observed from the aerial photography, whereas the value for mean bank retreat is generated for one bank only. Lateral channel movement determined from air photos is comparable to bank retreat values only if the channel position changes with no change in width. If during lateral channel movement bankfull width also increases as is commonly the case (i.e. one bank retreats faster than the other advances), the value of lateral channel movement will be lower than the bank retreat.

It should also be noted that no independent evaluation of the error associated with obtaining channel change data from historical maps is provided in Leeks et al., (1988). Such errors can be greater than those of photogrammetrically-derived data due to:

1. generalisation of medium-scale (i.e. 1:25,000 or 1:50,000) map features, notably the straightening of convoluted features (Mount et al., (2000).
2. The manual trigonometric re-survey methods used prior to 1948 which contain significantly greater error than modern techniques which utilise accurate photogrammetric and satellite positioning (Mount et al., 2000).

Table 8.1 Mean and maximum annual width change and lateral channel movement rates for the Trannon between 13 and 14 km from the source. Also shown (in bold type) is the mean historical bank retreat rate (obtained from historical maps) for a 200 m reach of the Trannon at Church Farm and presented in Leeks et al. (1988).

Year	Mean width change ($m\ yr^{-1}$)	Max. width change ($m\ yr^{-1}$)	Mean lateral movt. ($m\ yr^{-1}$) / **mean historical bank retreat**	Maximum lateral movt. ($m\ yr^{-1}$)
1885-1948			**≈ 0.19**	
1948-1976	0.01	0.62	0.13	0.78
1963-1976	0.19	1.68	0.37	1.88
1988-1995	0.02	1.96	0.39	3.04

Mean and maximum lowland channel change rates for the period 1948 – 1976 are at least half those for the period 1963 – 1976 indicating that the majority of lowland channel change occurred between 1963 - 1976. Indeed, this implication is further supported by the data presented in figures 6.9 and 6.11 in which the channel change observed between 1963 - 1976 is of a similar magnitude and pattern to that observed between 1948 – 1976. One may therefore deduce that the channel between 1948 - 1963 was relatively stable, with change rates lower than those observed for the period 1948-1976, and this deduction is confirmed when one visually compares the channel photographs from 1948, 1963 and 1976 in which only minor planform change is visible between 1948 and 1963 compared to major change between 1963 and 1976. (figure 8.1). This being the case, the rate of mean lateral movement between 1948 and 1963 was most probably similar to the mean rate of historical bank retreat for the preceding 63 years (1885 – 1948). However, this statement is made cautiously because the accuracy of Leeks et al.'s historical rate is unknown and probably considerably less than that of the photogrammetrically-derived results presented in this study. One may also deduce that in the period 1963 – 1976 channel change rates augmented significantly compared to the period 1948 – 1963 indicating increased channel instability in this period.

As discussed in chapter 2, changes in channel width and lateral stability can be linked to changes in the sediment (in particular bed load) and flow magnitude and frequency of the channel. Change of the magnitude observed on the lowland Trannon would appear to indicate either:

1. a significant increase in bed load supply in the period 1963-1976
2. a significant increase in the magnitude and frequency of competent discharges in the period 1963-1976
3. a combination of both.

Moreover, the timing of the response is between 15 and 28 years after the afforestation of the upland catchment. Consequently, the data implies that the destabilisation of the lowland channel might be a lagged response to the impacts of upland catchment forestry.

Between 1988 and 1995 the rate of width change and lateral movement increases further. This is most probably linked to the increased magnitude and frequency of competent flow in the period as demonstrated in section 5.3.

The above discussion is, however, incomplete without reference being made to the palaeochannels apparent in the floodplain in the 1963 aerial photography (figure 6.13). It is not possible to discuss the history of these palaeochannels quantitatively because their age is unknown, however they do serve to suggest that the current instability of the Trefeglwys reaches may not be without historical precedent. Indeed, it would appear that over the longer-term the Trefeglwys reaches have experienced significant positional change unrelated to upland land use suggesting that the stability of the lowland Trannon has been historically variable.

8.1.3 Lowland Trannon bed load transport rates

The augmentation of the lowland channel change rates is lagged between 15 and 28 years behind upland catchment afforestation. The Trefeglwys study reach is situated 7.5 km downstream of the point at which the Trannon exits the forest. Therefore, according to equation 8.1, forest-derived bed load, transported as a coherent pulse, would have to be transported to the Trefeglwys study reach at a mean rate of between 268 m yr^{-1} and 500 m yr^{-1} in order to arrive between 1963 and 1976 and hence be implicated as an important factor in the cause of increased rates of lowland channel change. This transport rate is high although not without example in the literature. Madej and Ozarki, 1996 (see figure 2.9) show a forest generated sediment wave translating a distance of 800 – 1600 m yr^{-1} in Redwood Creek, California. However, independent bed load transport measurements for the Trannon are required to determine whether a similar translation rate may have occurred in the middle Trannon.

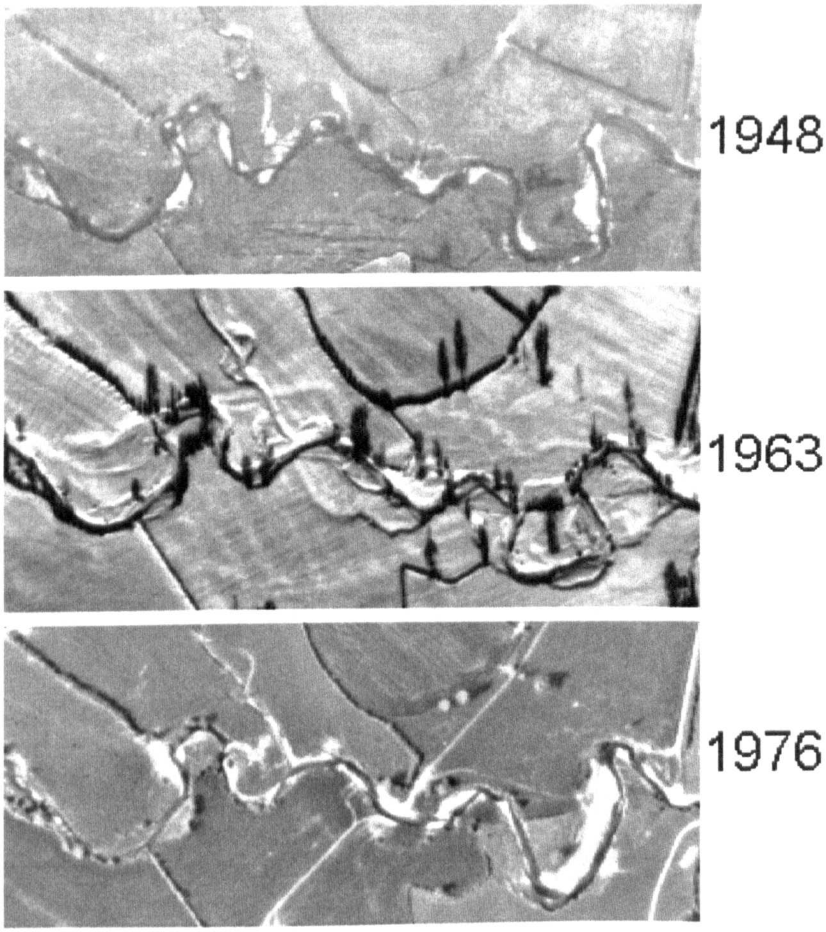


Figure 8.1 Non-rectified air photographs of the lowland Trannon at Trefeglwys for 1948, 1963 and 1976. Note the similarity of the planform during 1948 at 1963 and the evident instability between 1963 and 1976.

Historical bed load transport measurements are unavailable for the transfer reaches of the Trannon. Indeed, the only bed load transport data currently available for the Trannon are:

1. The morphologically derived bed load transport rate for the lowland channel derived for this study between October 1997 and February 2000 (see section 7.7.2)
2. A magnetic tracer experiment in the lowland reach by Leeks et al., (1988) for the period 08th February to 15th March 1982.

The bed load transport rates observed in chapter 7 at Bodiach Hall are presented in kg m^{-1} day^{-1} making them difficult to compare with the necessary mean bed load transport rate determined using equation 8.1 which is in m yr^{-1}. Leeks et al., (1988), however, provide some evidence that certain fractions of the bed load may be transported over large distances in a single event, even in the lowland channel. Their tracer results are presented in m yr^{-1} and are summarised in table 8.2 below:

Table 8.2 Magnetic susceptibility tracer results reported for the lowland Trannon at the upstream margin of the Trefeglwys study reach by Leeks et al. (1988).

Size (mm)	**% by weight**	**Maximum distance travelled over 35 day period (m)**
> 45	11.3	> 5 m
22.4	20.8	up to 123 m
11.4	20.3	up to 123 m
5.6	15.6	> 500 m
2.8	11.2	> 500 m
1.4	9.9	> 500 m
< 1.4	10.9	> 500 m

The results demonstrate that in the lowland channel the fractions of the bed < 5.6 mm were recorded up to a maximum of 500 m downstream of the seeding location after only 35 days in the channel with particles up to 22.4 mm being found up to a maximum of 123 m from the seeding location. One would expect that in the transfer reaches these rates would be even higher suggesting that the maximum transport distances of bed load < 22.4 mm may indeed be of the magnitude necessary to deliver forest-derived bed load to the lowland reaches between 15 and 28 years.

Whilst these data appear to support high transport distances of bed material in the Trannon they are misleading because:

1. they present maximum transport distances rather than the mean distances required in equation 8.1. The mean transport distance is not provided by Leeks et al. (1988) and it is likely that only a small proportion of each of the bed load fractions was transported close to the maximum distance.

2. the percentage of fine bed material (i.e. < 5.6 mm) reported in the bed of forested upland streams in mid-Wales is low and hence has limited significance regarding lowland channel change. Sawyer (2000) reports approximately 12 % of the bed load by mass in the Nant Tanllwyth to be < 5.6 mm. By comparison, approximately 78 % of the Tanllwyth bed is composed of material between 5.6 mm and 63 mm. The bed load in the Tanllwyth is derived from the same regolith as that of the neighbouring Trannon and similar forest ditch erosion processes are though thought to be responsible in both catchments. It would seem likely that the % of material < 5.6 mm is also small in the Trannon. As such, the mean transport rate of coarser bed load fractions is of greatest significance and Leeks et al., (1988) show this to be extremely low, at least in the lowland reaches.

Clearly, the determination of the transfer zone (see figure 1.1) bed load transport rates is a pressing further research need. Presently, it is not possible to determine a transfer time for the forest-derived bed load to reach the lowland reaches although the lack of any visible storage upstream of the Trefeglwys reaches would imply that bed load transport is rapid through the transfer reaches. The further investigation of bed load transport rates and the mechanisms of bed load transfer through the middle reaches of the Trannon is therefore identified as requiring further research.

8.1.4 Historical change in lowland bed load storage

Historical analysis of the extent of gravel bar area in the Trefeglwys reach has tentatively been presented in section 6.12 as an indicator of stored gravel volume. Such an analysis is not without comparison. Jacobson and Bobbitt Gran (1999) also used the extent of gravel bar area to infer change in the volume of stored gravel in the Ozarks, USA. In this study sub-aerial gravel bar area between 13 and 14 km from the Trannon source has been shown to increase by a factor of 1.4 between 1948 and 1976 (from 1.59 ha to 2.23 ha – figure 6.15). The timing and location of this increase corresponds with the timing and location of increased rates of channel widening and lateral channel movement. It would, therefore, appear to support the view that destabilisation of the Trefeglwys reaches can be linked to an increase in sediment storage. However, the data presented lack a stage record for the 1948 aerial imagery and therefore does not permit an assessment of the impact of stage on the gravel bar area at 1948 to be made. Moreover, the impact of channel morphology change on the position of the water line on each photograph is also unknown. Consequently, it is the author's view that the increase in gravel area between 1948 and 1976 is not of a sufficient magnitude to categorically demonstrate an increase in the volume of bed load stored in lowland gravel bars.

There is an indirect line of evidence which provides further evidence that lowland reach destabilisation is linked to deposition of bed load onto bar forms. As previously noted, the downstream pattern of lateral instability in the Trefeglwys reach appears to match well the location of point bars. This is especially true for lateral change between 1948 and 1976 where two localised zones of lateral instability exist at 13.6 – 13.8 km and 14.5 – 14.7 km (figure 6.11 and 6.12). The position of these zones concurs with the location of meander bends and point bars visible on the aerial photography. Hence the lateral change in these locations is most likely a result of bed load deposition onto point bars, subsequent progradation of the bars, flow diversion and outer bank erosion. Such a mechanism is subsequently shown in chapter 7 to be the main mechanism of contemporary channel change.

8.1.5 Downstream change in stream power – a feasible cumulative effect

As presented in section 2.9 sediment accumulation within a channel network has been recognised as being a result of:

- The character of the channel network and the location and timing of exogenous sediment yielded to the tributaries. Essentially the cumulative effect is a result of elevated tributary sediment yields and the nature of its sediment routing through the tributaries (i.e. Jacobson and Bobbitt Gran (1999)).
- Localised inputs of sediment to the main channel from tributaries (i.e. Church (1983)).

In this catchment, in which every tributary has been walked in the field and examined in respect of any likely sources of elevated bed load supply, the majority of bed load yielded to the lowland channel is thought to be associated with Dolgau forest and the upper catchment. None of the tributaries displayed any visible evidence of recent extensive erosion or land use changes which may have elevated their bed load inputs to the main channel. Consequently, it would seem unlikely that timing and routing of bed load inputs to the main channel from tributaries can be implicated as providing a mechanism by which bed load accumulation results in increased rates of channel change in the lowland reaches. Instead, the current location of the upstream most major bed load storage sites, which coincide with the beginning of the destabilised reaches at Trefeglwys, would suggest that a decrease in the bed load transport rates at Trefeglwys may exist and be of significance.

- A decline in the bed load transport rate.

In the absence of bed load transport data for the channel between the forested upper catchment and the destabilised lowland reaches it has not been possible to determine if a drop in bed load transport rates does exist at Trefeglwys through direct field quantification of bed material transport rates and this is highlighted as an area which would benefit from further study. However, downstream variability in gross and unit stream power has been examined (section 7.8).

Unit stream power is known to broadly relate to bed load transport rates (Bagnold, 1980; Gomez and Church, 1989) to the extent that in the context of a common bed load size fraction, in regions of high unit stream power one may expect high rates of bed load transport, and conversely in regions of low unit stream power one may expect deposition. Consequently, examination of the spatial nature of variability in unit stream power, particularly any decreases, may be used to imply the spatial variability of bed load transport rates.

Downstream variation in both gross and unit stream power, together with the locations of the upper catchment forestry and stable and destabilised reaches are shown in figure 8.2. The shape of the gross stream power plot appears to conform well to that predicted by Lawler (1992) in figure 2.10 for Welsh rivers, with a mid-river peak occurring at approximately 8 km. According to Lawler (1992) a mid-river peak in gross stream power is caused when:

1. discharge downstream increases as a power function of distance:

$$Q = kL^m$$

(equation 8.2)

where

Q is discharge (cumecs)

k is a constant

L is distance from source (km)

m is a power function

2. Slope downstream declines exponentially according to the formula:

$$S = S_0 e^{-rL}$$

(equation 8.3)

where

S is the channel gradient at any point downstream (m m^{-1})

S_0 is initial slope at some upstream reference section (m m^{-1})

L is the distance from source (km)

r is the coefficient of slope reduction

When plotted, it can be seen that the estimated bankfull discharge does fit the power function model reasonably well (figure 8.3), however, estimated bankfull discharges at both 8.2 km and 10.6 km are higher than the best fit power function line predicts. The relationship between slope and distance from the source though does not conform to the simple exponential model because the upper 4 km of the Trannon cross the Cambrian plateau, a relatively flat upland landscape where channel gradients seldom exceed 0.01. This means that the downstream change in slope is sigmoidal rather than exponential (figure 8.4). The pattern of downstream change in channel slope is much better described by a third order polynomial trend line than a negative exponential. Consequently, the mid-channel peak in gross stream power at 8 km is not, as Lawler suggests, a result of a combination of a negative exponential slope function and a positive discharge power function. Rather, the peak occurs because of a combination of high channel gradient (relative to that of the upland plateau) and relatively high estimated bankfull discharge (probably caused by the

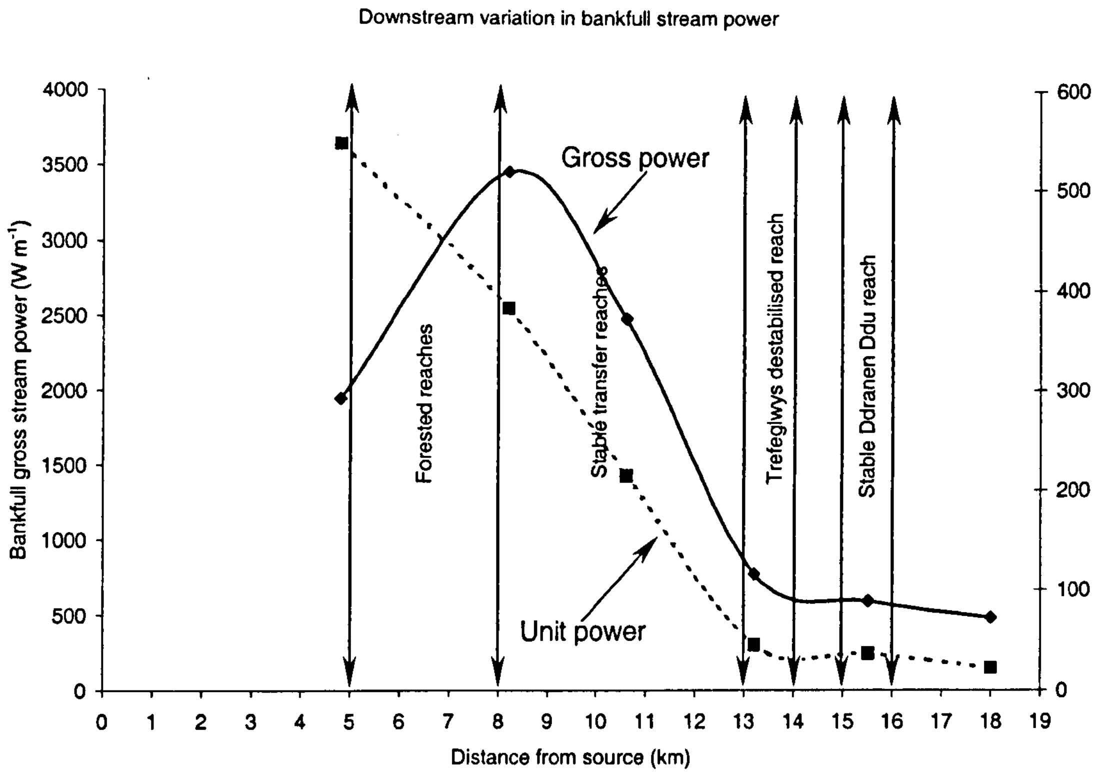


Figure 8.2 Downstream variation in unit and gross stream power with locations of forestry, stable transfer, unstable Trefeglwys and stable Ddranen Ddu reaches shown.

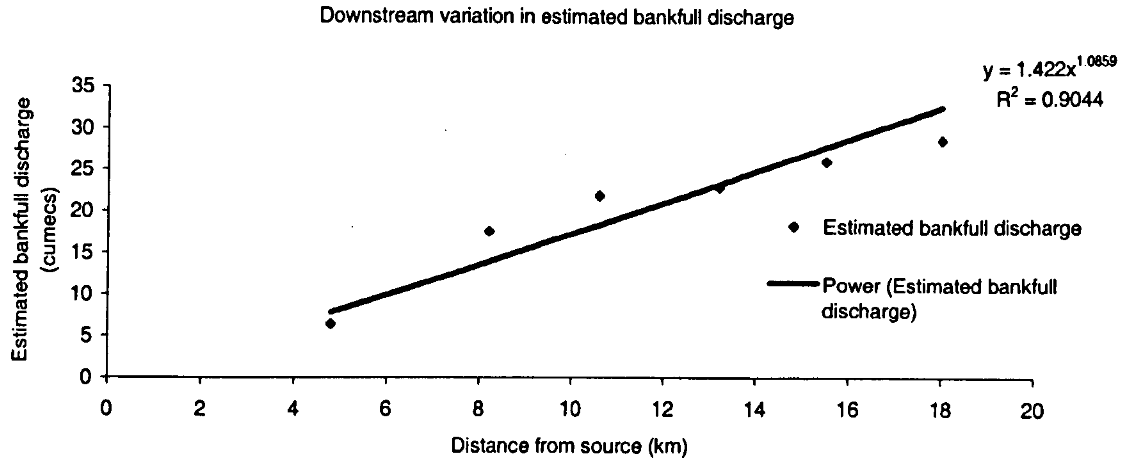


Figure 8.3 Downstream change in bankfull discharge with power function relationship fitted. Note that the power function relationship is almost linear.

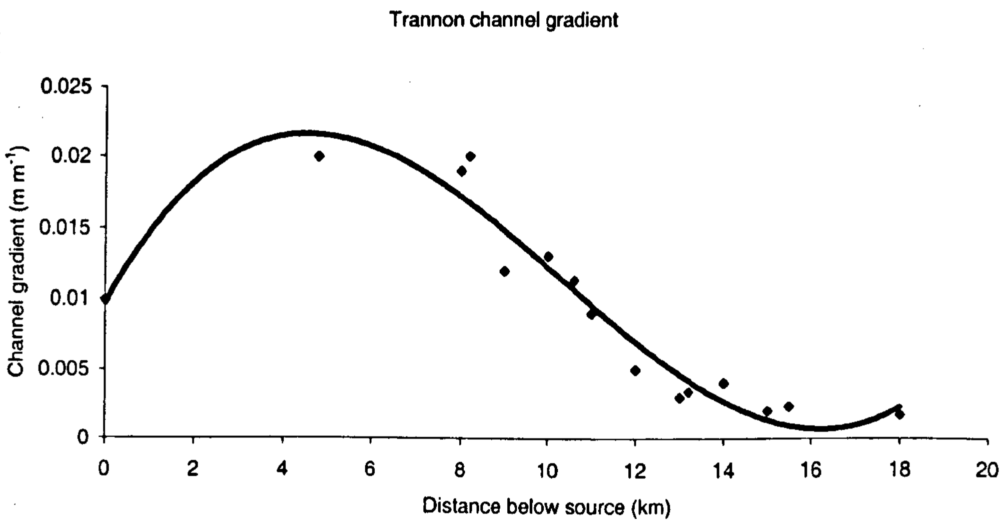


Figure 8.4 Downstream change in channel gradient determined from the 1:25,000 scale O.S. map.

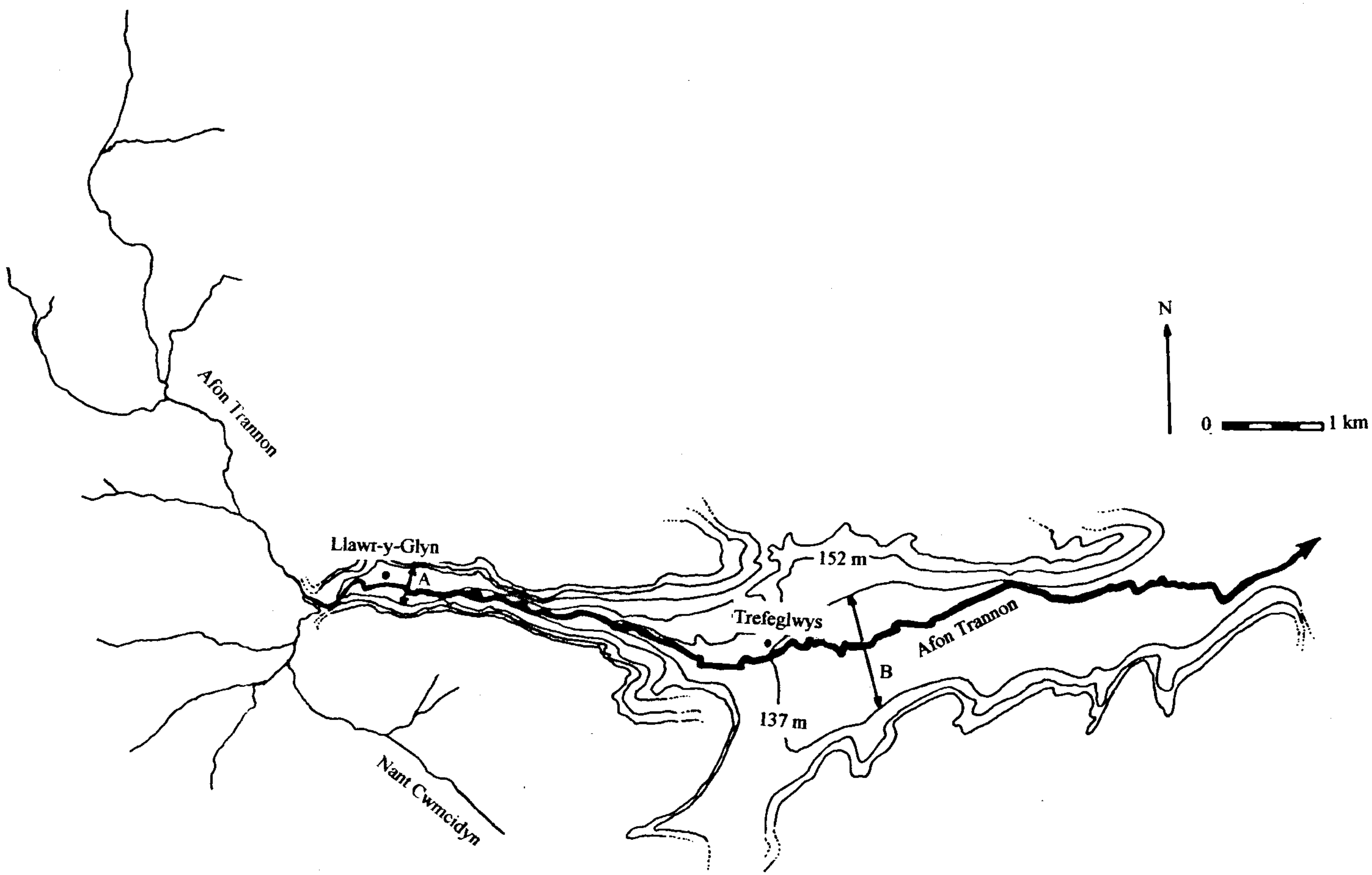


Figure 8.5 Sketch map of the Afon Trannon with middle valley and lowland valley contours added. The confined middle valley is labelled A and the broad lowland trough is labelled B.

upstream confluence of a large right hand tributary called the Nant Cwmcidyn – see figure 8.5) at this point.

A lack of conformation between the Trannon and the Lawler model is also evident in the downstream change in unit stream power. Unit stream power does not peak in the middle reaches with gross stream power as predicted by Lawler, rather it declines quickly, and at a fairly consistent rate, between 4.8 and 13.2 km before stabilising at, and changing relatively little in, the lowland reaches. The relatively wide and shallow channel at 8.2 km, compared to the narrow incised channel above the forested reach (caused by down cutting into the upland peat), means that unit stream power is lower in the middle reaches of the Trannon than in the upper most reaches, hence the peak in gross stream power is not mirrored in the unit stream power data. It is the downstream pattern of unit stream power, rather than gross stream power that is of greater significance for bed load transport rates (see equation 2.1) and hence cumulative effects.

The downstream pattern of unit stream power in the Trannon is one which could conceivably cause the accumulation of bed load in the Trefeglwys reaches. In order for bed load transport rates to decline in a downstream direction a steep drop in unit stream power is required. A shallow decline would have limited impact because bed load size commonly decreases in a downstream direction, therefore a lower unit stream power is required to transport bed load with increasing distance downstream. Figure 8.2 shows that in the Trannon there is a consistent, extremely steep downstream decline in unit stream power (67.4 W m^{-2} km^{-1} between 8.2 km and 13.2 km). Such a steep energy gradient implies that bed load at a point in the channel upstream of 13.2 km would be moving slower than bed load of the same size at another point further upstream and that the effect of decreasing bed load size downstream would be minor. Consequently, bed load in the Trannon can be thought of as continuously accumulating with transport downstream. If a mechanism exists in which this constantly accumulating bed load enters storage, rather than continuing its downstream transport, bar growth and lateral instability may occur.

Such a mechanism exists at Trefeglwys where the Trannon leaves the relatively confined valley of the middle, transfer reaches. In the middle reaches the floodplain width ranges from approximately 150 m to 375 m and bedrock is still occasionally visible in the Trannon bed (Leeks et al., 1988, pg. 208). Below this the river enters the broad glacially scoured trough of the lowland reaches in which the floodplain width is commonly in excess of 2000 m and drift deposits have a depth of up to 68 m (figure 8.5). This shift in valley character is marked by a decrease in the valley gradient (0.02 at 8.2 km compared to 0.0034 at 13.2 km) and a decrease in the downstream increase in channel bankfull width (1.05 m km^{-1} between 8.2 and 10.6 km compared to 0.87 m km^{-1} between 13.2 and 18 km). This has the effect of greatly reducing the downstream rate of reduction in unit stream power below 13.2 km, where it ranges between 45.47 W m^{-2} at 13.2 km and 22.78 W m^{-2} at 18 km. This range of bankfull stream power is extremely significant as it falls within the range, observed by Nanson and Croke (1992), at which the major mechanism of channel adjustment changes from being mainly vertical to lateral (see table 8.3). It is also close to the median stream power of active meandering channels observed by Ferguson (1981,1987) and similarly the range of stream power between 8 km and 13 km in the Trannon fits the range of confined channels observed by Ferguson (1981, 1987) and presented in table 8.4.

At 13 km the valley character changes sharply, from that in which floodplain width is constrained and bedrock is overlain by only thin deposits to that in which floodplain width increases enormously and the thickness of floodplain deposits is commonly in excess of 50 m. This promotes the change in channel character from a channel in which lateral adjustment is restricted by only thin floodplain deposits to one in which bank erosion and hence the deposition of lateral bars occurs meaning lateral channel adjustment is possible. This then forms the upstream most site for significant deposition of bed load generated in the upper catchment. It also explains why the channel displays a high level of lateral stability at Llawr-y-Glyn where the low bank heights (mean = 0.49 m at Llawr-y-Glyn compared to 1.09 m at

Trefeglwys) demonstrate the thinness of floodplain deposits and hence the still significant influence of bedrock on channel morphology.

Table 8.3 The relationship between floodplain / valley character, channel character, unit stream power and bed load transport and deposition. After Nanson and Croke, (1992).

Floodplain type / valley character	Channel character / bankfull unit stream power	Bed load transport / deposition
Confined floodplain (high energy)	• Relatively narrow channel • High gradient • $\omega = 300 - 1000$ W m^{-2}	• High bed load transport rates • Lateral deposition limited • Channel adjustment is mainly vertical
Non-confined, scrolled floodplain (medium energy)	• Relatively wide meandering channel • Non-cohesive floodplains • Low gradient • $\omega = 10 - 60$ W m^{-2}	• Low bed load transport rates • Sediment deposited as accretion onto lateral point bars • Channel adjustment is mainly lateral

Table 8.4 Bankfull stream power of British rivers. After Ferguson, (1987).

Channel Pattern	No. of rivers	Range of ω	Median
Inactive unconfined	25	1-60	15
Confined	24	20-350	100
Active meandering	40	5-350	30
Active low sinuosity	6	120-300	160

8.1.5.1 Cumulative effects – summary

The above discussion serves to demonstrate that although a mid-channel peak in unit stream power does not occur in the Trannon, and the reason for a mid-channel peak in gross stream power does not conform to the Lawler model, a mechanism does exist by which sediment is able to accumulate at Trefeglwys. The accumulation is a result of the combination of consistently declining unit stream power which acts to progressively consolidate the non-point source bed load inputs of the upland catchment between 8 and 13 km from the Trannon source coupled with significant change in the valley character at 13 km which means lateral bed load deposition becomes possible.

8.1.6 Medium-term change in flood magnitude and frequency

The modelled daily mean stage (DMS) record, presented in chapter 5, allows assessment of change in flow magnitude and frequency to be made for the period 1969-1999. A lack of rainfall data for the period 1948-1968 means that flow magnitude and frequency prior to 1969 cannot be assessed. The main features of the DMS record are presented in table 8.5.

Table 8.5 Comparison of the variability in DMS between 1969-1988 and 1989-1999.

Parameter	**1969-1988**	**1989-1999**
Frequency (per annum) with which DMS > 0.89 m	Mean = 10.23 days Range = 23 days Standard deviation = 6.06 days	Mean = 24.27 days Range = 57 days Standard deviation = 16.95 days
Maximum DMS attained in entire period	Maximum = 1.5 m Range = 0.77 m	Maximum = 2.23 m Range = 0.97 m

Between 1969 and 1988 the frequency with which DMS exceeded the entrainment threshold, is shown to be low with a small range and modest standard deviation. This is in contrast to the situation for 1989 and 1999 when the frequency doubles and the range and standard deviation are inflated. This relative stability between 1969 and 1989 is also borne out by the maximum DMS data. The cause of this increase in the magnitude and frequency of flows since 1989 has not been established in this study, but would appear to be a result of change in the antecedent moisture conditions in the catchment, possibly resulting from altered temporal patterns of rainfall in the last ten years. However, to establish the cause would require further, substantive study outside of the remit of this work. Consequently, it forms an interesting issue for future research.

The DMS records presented in Chapter 5 suggest that channel change measured between 1963 and 1976 photographs is probably not a result of substantive flow pattern change, and that sediment loads to the lowland channel have been fundamental to the destabilisation of the channel in the period. However, there is a large question mark over the initial six years between 1963 and 1969 where no data are available. Between the 1988 and 1995 photography flow magnitude and frequency appear to have altered substantially, with increased frequency and magnitude of competent events. Such increases might be expected to cause channel incision and widening, and indeed the mean channel width does increase at Trefeglwys in this period but this increase is small compared to the increased lateral instability in the period. It would seem likely that the increased laterally instability between 1988 and 1995 is related, to some extent, to the increase in competent flow magnitude and frequency rather than solely being a result of increased bed load yields, however, determining the relative importance of each is extremely difficult when the temporal resolution of the channel change data is multi-year as in this project.

8.1.7 Evidence of an upstream bed load deficit

A fully quantified, medium-term sediment budget for the lowland Trannon is not possible to construct due to the limited archive data available in this project, in particular the lack of bed load transport rate and storage volume data. However, an observation can be made which strongly suggests that the historical upstream inputs to the Trefeglwys reaches fell short of those required to produce the observed patterns of channel change.

The upland catchment bed load yield estimates suggest that if forest bed load yield is to be implicated as the direct cause of the observed lowland instability between 1963 and 1976, a maximum of 264 t of upland catchment bed load is responsible for the lowland channel change (assuming a 15 year transport lag time). This value is generated from the estimated upland catchment bed load yield between 1948 and 1961 presented in figure 4.6. Converted to a volume the upland catchment bed load input represents 99.62 m^3

This addition can be placed into context if one assumes that all of the upland catchment bed load enters storage in the lowland channel and is distributed evenly onto the sub-aerial gravel bar surfaces present in the lowland Trannon in 1948 (15870 m^2 – figure 6.15). It would represent an insignificant mean increase in bar surface elevation of just 6 mm and hence it is extremely unlikely that it had the potential to drive significant, large-scale channel change.

8.1.8 Summary

The lines of evidence necessary to implicate upland catchment forestry as the direct cause of medium-term lowland channel instability were identified in section 8.1.1. These are restated in table 8.6 together with the main findings of the preceding discussion. Of most significance is the extremely limited impact of the relatively small bed load yields of the upper Trannon on bar elevation (point 5). Indeed, this evidence demonstrates conclusively that there was simply not enough bed load released from the upland catchment to directly generate the increased bed load storage and hence rates of channel instability observed between 1963 and 1976. Consequently, a further bed load source is required to explain the patterns in the historical data and this has been attempted by examining the contemporary sediment budget of the lowland channel.

Table 8.6 Summary of evidence for a direct link between elevated upland catchment bed load yields and lowland channel instability.

Line of evidence	Research findings	Is the evidence for or against a direct link between upland afforestation and lowland channel instability?
1. Timing of lowland channel instability (i.e. 1963-1976) should lag the timing of afforestation by the mean bed load transport rate multiplied by the length of the transfer zone.	No independent analysis of bed load transport rates in the middle Trannon reaches makes definitive statement difficult. Leeks et al., (1988) trace results suggest fine bed load fractions may be transported to the lowland reaches in the observed lag time, however Sawyer (2000) shows the fine fraction to be a small proportion (< 20%) of the overall bed load in the neighbouring Tanllwyth and it is likely to be of similar significance in the Trannon.	Inconclusive – requires further research
2. Historical bed load storage should be seen to increase in the reach experiencing channel instability for elevated bed load to be implicated. The timing of increased storage should match the timing of instability.	Gravel bar areas suggest this to be the case, however uncertainty in the error associated with bar area means that the link cannot be categorically stated. However, historical patterns of lateral movement do imply increased storage of bed load in barforms between 1948 and 1976.	For

Line of evidence	**Research findings**	**Is the evidence for or against a direct link between upland afforestation and lowland channel instability?**
3. A lowland bed load accumulation mechanism should be evident.	The downstream pattern in unit stream power provides a mechanism whereby bed load would progressively accumulate during its transport thorough the middle Trannon.	For
4. It should not be possible to explain the augmentation of lowland channel change rates solely by changing patterns of streamflow.	Whilst variability in the annual patterns of competent flow magnitude and frequency are low in the period 1969-1976 no data is available for the period 1948 – 1963. Consequently, it is not possible to discount the impact of altered streamflow patterns on the period prior to 1969.	Inconclusive – required historical data are not available
5. The quantity of bed load yielded from the upland catchment should approximately equal that entering storage in the lowland reaches	When the historical bed load yield volume is applied to the exposed gravel bar area, the increase in bar elevation is negligible. It therefore seems impossible that historical forest bed load yields alone could have produced the observed channel instability.	**Strongly against**

8.2 Lowland bed load additions – the sediment budget

The sediment budget equation presented in equation 7.1 may be amended thus:

$$U_I + L_I = \Delta S + O$$

(equation 8.4)

where

U_I = upstream bed load inputs (i.e. upland catchment, upstream tributaries, upstream banks)

L_I = local inputs

ΔS = change in bed load storage

O = bed load outputs

Local inputs to reach-scale sediment budgets commonly result from the erosion of bank faces. Local bank material inputs have been recognised in several studies as providing significant additions to lowland sediment budgets (table 8.7) and it is feasible that they form a major addition to the lowland sediment budget of the Trannon.

Table 8.7 Examples of % inputs to sediment budgets from river banks.

Author / date	**% input to budget from bank erosion**
Batalla et al. (1995)	34 %
Duysings (1986)	53 %
Peart and Walling (1988)	30 %
Barker (2000)	20 %

8.2.1 Lowland Trannon local bank inputs

Accompanying the transition from a confined valley to a broad, low gradient trough above Trefeglwys is a transition from stable, vegetated sand, silt and clay banks of low height to composite gravel banks commonly in excess of 1.0 m in height and ranging up to 2.5 m high. Visual inspection of the banks has provided an estimate of the percentage gravel (8 mm or larger) in these lowland banks of up to 50 %, and a mean gravel content of 16.1 % (section 7.9) whilst calibration of the observations via a bulk sample suggests that the visually derived data is an underestimate of the true content (section 7.9.2) by as much as 18 %. Moreover, the analysis also showed that a significant proportion of the bank (30 %) was composed of material between 1mm and 8 mm in size which would also contribute to channel bed load. This material was not quantified by visual bank mapping. Consequently, bank erosion in the Trefeglwys reaches has the potential to provide significant additional bed load to the lowland channel.

In the following budget calculations it is assumed that bed load from the upland catchment forestry has a transfer time of 15 years to the lowland reach (i.e. it is assumed that upland catchment bed load did reach the destabilised lowland channel reaches by 1963).

8.2.2 Medium-term bed load addition from lowland bank erosion

The medium-term additions of bed load > 8 mm from lowland bank erosion (section 7.9) is compared to that from the upland catchment (Chapter 4) in table 8.8 below.

Table 8.8 A comparison of the medium-term bed load yield from the upland catchment and lowland banks at Trefeglwys

Input	**Thesis section value obtained from**	**Value as volume (m^3)**	**Value as mass (t)**
Gravel >8 mm additions from bank erosion at Trefeglwys reaches (1948-1995)	Section 7.9 Table 7.15	3117.9 m^3 Min: 437.2 m^3 Max: 5798.5 m^3	5783.7 t Min: 811 t Max: 10756.2 t
Gravel > 8 mm additions from upland catchment (1948-1980) i.e. 15 year transport time.	Figure 4.5 and section 7.9.2.2. (Sawyer 2000 bed load size results)	783.2 m^3	1453 t

The potential of bank erosion to supply large volumes of bed load to the lowland Trannon is well demonstrated. The data suggest that the total medium-term erosion of the composite lowland banks at Trefeglwys contributed 5783.7 t of bed load > 8 mm direct to the lowland reaches although significant error in the aerial photography analysis (table 6.7) means that the figure could be as low as 811 t or as high as 10756.2 t. This is in contrast to the total estimated upland catchment bed load yield (of > 8 mm) to the Trefeglwys study reaches of 1453t, although as previously mentioned in chapter 4, this figure is thought to be rather high. Even if the lowest bank yield estimate is employed, the above data demonstrates that coarse bed load additions from the lowland banks is at least of an equal order of magnitude to that of the coarse bed load yield from the upper catchment, and probably much greater.

The above analysis is restricted in that it only examines the coarse bed load fractions (i.e. material > 8 mm) with no analysis of the importance of finer fractions. This is a restriction of the visual mapping procedures employed in section 7.9. In the following discussion of the contemporary sediment budget all bed load material sizes (i.e. > 1 mm) are considered because the local input data are unrestricted by a visual bank material mapping procedure. Rather, local bank inputs are derived volumetrically from DTM data, and converted to a mass, in which all grain sizes > 1mm are included.

8.2.3 Contemporary bed load additions from banks

To investigate the lowland bank material inputs further, with the inclusion of all bed load size fractions > 1mm, the contemporary sediment budget of the Bodiach Hall reach has been investigated. The budget is constructed for the entire 830 day field study period at Bodiach Hall (i.e. 31.10.1997 – 08.02.2000) The importance of local bed load additions to the contemporary Bodiach Hall reach-scale sediment budget can be well demonstrated using a combination of data from the channel DTMs and the bank bulk sample data removed from the bank opposite PB3, the major local input of bed load to the study reach (Chapter 7).

Differenced DTM analysis of Bodiach Hall study reach (figure 7.10 a-c) demonstrated that the bank opposite PB3 was the main location for bank erosion in the contemporary field study. At this location the volume of bank erosion matched, almost exactly, the volume of gravel deposited on the point bar (figure 7.21). The linear relationship between deposition and erosion at the site is:

$$y = 0.9119x \qquad \text{(equation 8.5)}$$

where

y = bank erosion (t)

x = deposition of gravel onto PB3 (t)

The bank bulk sample from the outer bank shows 88 % of the bank is composed of material of bed load size (> 1mm) (see table 7.13). This percentage can be substituted into the above equation where:

- x = deposition of bed load onto PB3 which is equal to 484.93 t ± 69.26 t
- y = 88 % of the volume of bed load entering the channel as a result of bank erosion opposite PB3 over the contemporary study period, which equals 389.14 t ± 60.94.

The bed load addition from the bank opposite PB3 can be compared to the upland catchment yield over the 830 day contemporary study period. If a transfer period of 15 years is again assumed the material entering the lowland channel between October 1997 and February 2000 will have been yielded from the upland catchment between October 1982 and February 1985. Estimated annual bed load yields from the upland catchment have remained stable at 68.58 t yr^{-1} (section 4.3) since the forested area stopped growing in 1978. Consequently, if it assumed that all of the upland catchment bed load yield is stored in the Bodiach Hall study reach, the upland catchment bed load inputs to the reach (U_{ic}) may be calculated:

$$U_{ic} = \left(\frac{68.58}{365}\right) \times 830 = 155.94$$

(equation 8.6)

Consequently, it can be seen that the contemporary bed load addition from the bank erosion opposite PB3 (389.14 t ± 60.94) is more than double that of the entire estimated upland catchment bed load yield (155.94 t) in the same period.

However, the large bed load inputs from the bank erosion at PB3 present a budget problem downstream. The problem arises from the fact that the bed load additions from opposite PB3 are neither:

1. expressed downstream in the DTMs as deposition either as a translating wave or as deposition onto PB4
2. likely to have been transported out of the Bodiach Hall reach due to low bed load transport rates of only 320 kg day^{-1} above which flow was competent (i.e. > 0.89 m) (section 7.7.2). Over the 830 day study period there were 120 days where DMS > 0.89 m. Therefore the total bed load transport from the reach was just 38.4 t.

Barker (2000) was able to demonstrate via tracer experiments that in the meandering River Yarty, Devon, the majority of coarse sediment additions to the channel from bank erosion were transported downstream to the next meander point bar upon which they were deposited. It would seem feasible that such a process might also occur in the Trannon in which case one would expect the growth of PB4 to approximately equal the volume of gravel entering the channel at PB3. However, this is not evident in the Bodaich Hall DTMs. Indeed the addition of bed load from opposite PB3 is met with a decrease in the volume of PB4 of –24. t ± 133 t (see figure 7.20).

The above discrepancy is simply explained by the impact of regular bar dredging of PB4 between 1997 and 2000 (figures 8.6 and 8.7) to provide gravel for the construction of hard standing surfaces and track repairs at local farms. The bar is known to have been dredged immediately prior to survey 1 at Bodiach Hall (29 October 1997), prior to survey 4 (May 1999) and immediately after survey 5 (9-11 November 1999) because the dredging coincided with a field visit, however it is very possible that extraction also occurred at PB4 on other occasions.

Dredging on the 29.10.1997 was particularly deep, and left a well defined impression on the bar surface which is clearly visible in the cross profile surveys. Consequently, reconstruction of the bar surface was possible by extrapolating a gently curved surface between the margins of the dredged region on each of the affected cross profiles (BH23 – BH33). Figure 8.8 shows an example of the bar

Figure 8.6 Bar dredging operations at PB4.

Figure 8.7 PB4 after dredging showing the extent of the gravel extraction.

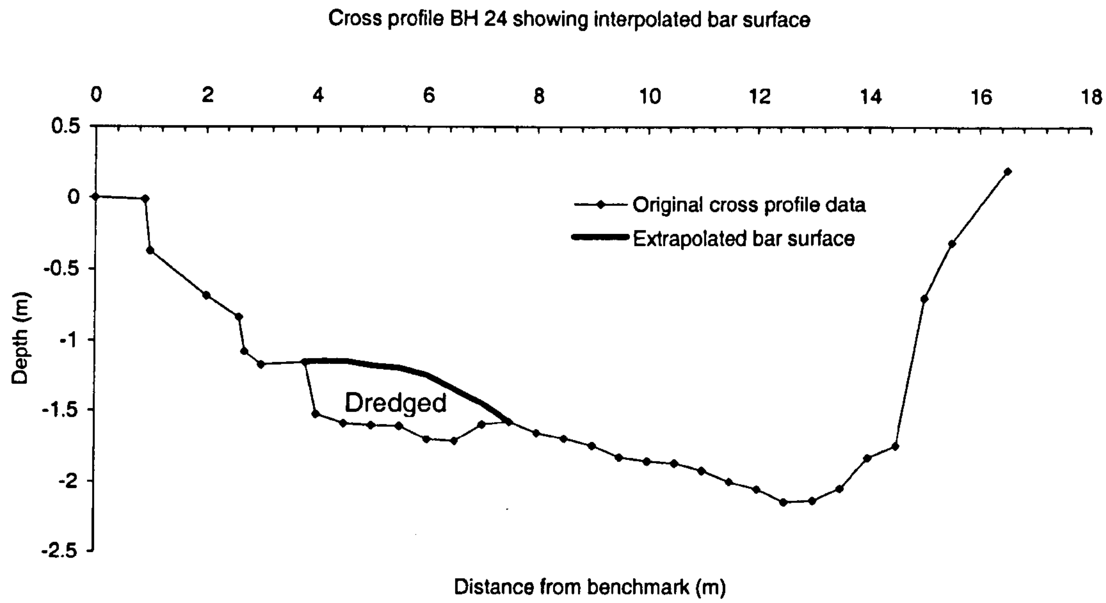


Figure 8.8 Example of bar surface interpolation used to estimate the volume of dredged material from PB4.

surface extrapolation from cross profile BH24. The volume of the reconstructed channel was determined using the TIN approach previously presented. Bar surface reconstruction shows that approximately 128 t of bed load was removed from PB4 on the 29.10.1997, equivalent to approximately 33 % of the total bed load input from the banks opposite PB3 over the entire study period. The extraction volume demonstrates that gravel bar dredging has had a large impact on the reach sediment budget of the Bodiach Hall reach and is the most likely reason for the identified gravel deficit and the overall decrease in the volume of PB4 over the field study period.

8.2.4 The budget diagram

A simple sediment budget diagram for the period 31.10.1997 to 08.02.2000, in which all of the upland catchment bed load yield for the period are assumed to have been deposited in the Bodiach Hall reach, and in which other upstream inputs are unaccounted for, is presented in figure 8.9. The dominance of lowland bank erosion inputs is clear with them providing 2.5 times the upland catchment bed load input in the contemporary budget. Dredging is also a potentially important factor but is restricted to the downstream margin of the study reach and is therefore not a major uncertainty in the sediment budget for most of the lowland channel.

8.3 The influence of discharge on contemporary and historical channel response

The maximum contemporary rate of lowland channel change measured at Trefeglwys is presented in table 8.9 together with the medium-term maximum rates as determined from aerial photography and reported in Leeks et al. (1988).

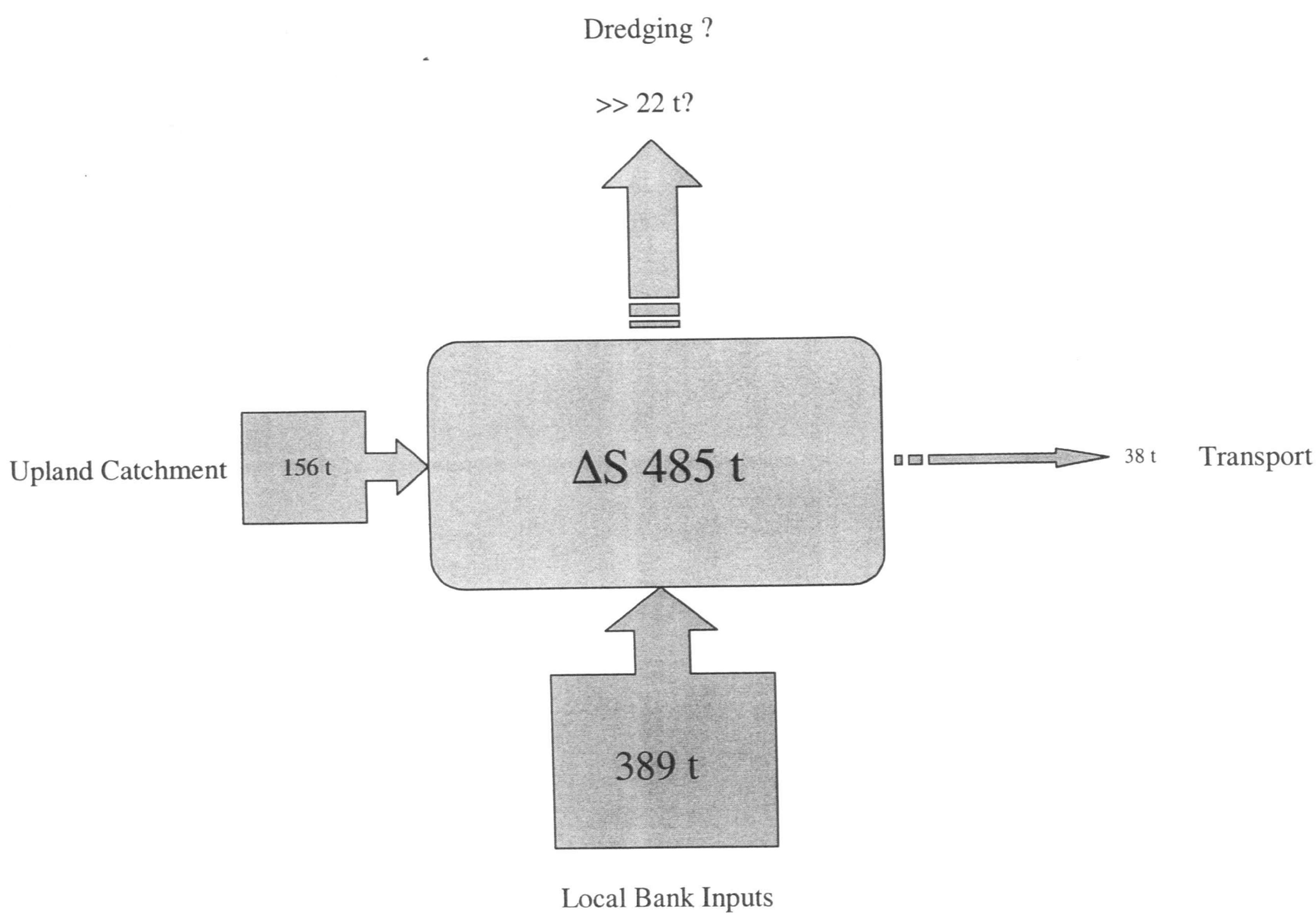


Figure 8.9 Simplified contemporary sediment budget diagram for the Bodiach Hall study reach 30.10.97 – 08.02.2000.

Table 8.9 Maximum contemporary and medium-term bank erosion rates at the lowland Trannon 1963 – 2000

Measurement	Rate of change (m yr^{-1})
Maximum lateral movement from aerial photography 1963 - 1976	2.04
Maximum bank retreat rate from Leeks et al. (1988) 1981 – 1983	0.96
Maximum lateral movement from aerial photography 1988 - 1995	3.04
Maximum contemporary bank retreat rate from repeat cross profiles October 1997 – February 2000	5.30

The table demonstrates a relatively high rate of change during 1963-1976, commensurate with the destabilisation of the river channel. In the period 1981-1983 the maximum rate halves. This is most likely due to the impact of the recent river engineering of the channel which, only four years after completion, probably retained some integrity. In the period 1988-1995 maximum rates increase to almost 1.5 times that in the initial destabilisation, and contemporary field data show that the increase in maximum rates has continued throughout the 1990s. Contemporary surveys demonstrate a maximum retreat rate more than 2.5 times greater than that experienced in the initial period of post-forestry destabilisation.

The cause of the extremely high rates of maximum change in the period 1988-2000 appears linked to the increased magnitude and frequency of competent discharges in the period rather than solely the deposition of sediment from upstream or from local banks. The link is well illustrated in figures 8.10 and 8.11 in which the contemporary bank retreat rates at the most active cross profiles in the BH study reach are compared to the maximum daily mean stage (DMS) and number of flow events exceeding the bed entrainment threshold. An apparently strong, yet

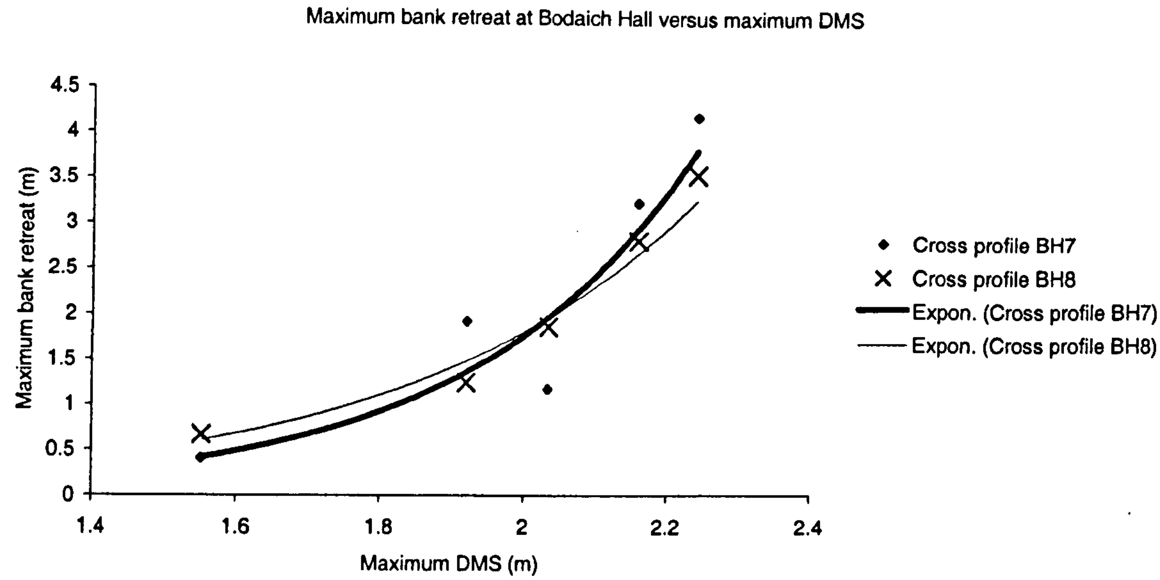


Figure 8.10 Maximum bank retreat versus maximum DMS in the Bodiach Hall study reach

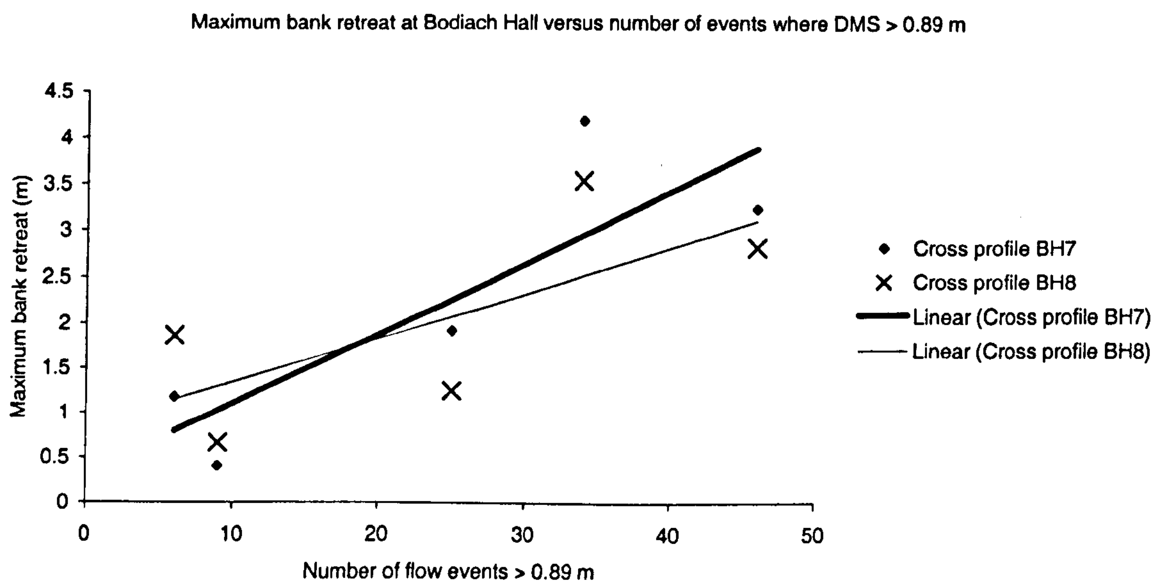


Figure 8.11 Maximum bank retreat versus number of flow events above the bed entrainment threshold.

statistically untested, exponential relationship exists between the maximum DMS and the total bank retreat in each survey period. Similarly, a positive relationship exists between the number of flow events > 0.89 m stage and the total erosion, however the relationship is less well defined with much greater residual values.

The relationships between the maximum DMS, the number of DMS events > 0.89 m and maximum bank retreat rates is not apparent throughout the historical data. Table 8.10 presents historical maximum bank erosion rates and, where possible, the maximum DMS attained, and the number of events in excess of 0.89 m stage during each inter-survey period.

Table 8.10 Maximum historical bank retreat rates and maximum DMS and number of events > 0.89 m stage in each inter-survey period.

Survey period	**Max. bank retreat rate ($m\ yr^{-1}$)**	**Maximum DMS in inter-survey period (m)**	**No. events > 0.89 m stage in inter-survey period / years ($No.\ yr^{-1}$)**
1963-1976	2.04	1.47	> 10.3
		(1969-1976 only)	(1969-1976 only)
1981-1983	0.96	1.44	14.3
(from Leeks et al., 1988)			
1988-1995	3.04	1.85	16.2
1997-2000	5.30	2.24	52.8
(field survey data)			

From the table it can be seen that in periods when both maximum daily mean stage and the number of flow events > 0.89 m are high (i.e. 1988 – 1995 and 1997 – 2000) the maximum bank retreat rates are also high. Conversely, where flow magnitude and frequency are lower (i.e. 1981 – 1983) maximum bank retreat rates are accordingly lower. Of particular interest is the period 1963 – 1976 in which maximum bank retreat rates are more than double those during 1981 – 1983, however the maximum daily mean stage attained is comparable for the two periods. This could be a result of either:

- The fact that the stage record only covers 1969 to 1976 and it is possible that in the period 1963 – 1969 the maximum daily mean stage was in excess of that in the period for which records exist.
- Deposition of bed load from upstream in the study reach causing elevated bank erosion rates despite only moderate maximum daily mean stage flow events.

In the absence of a longer flow record, the length of which is restricted by the lack of rainfall data in the study region prior to 1969, it is not possible to determine which of the above causes holds the most merit.

8.3.1 Peak discharge and upland catchment forestry – implications for lowland stability

The clear relationship between the magnitude of peak discharges and lowland bank erosion at PB3, and the speed of lowland channel response, suggests that the elevated peak discharges recognised during the felling and planting phases of upland catchment forests (i.e. Robinson, 1980) may have an important influence on lowland channel stability. It is not possible to determine whether the increase in peak discharge in the lowland Trannon equals the 40 % observed in the Coalburn catchment, in which the entire catchment was ploughed and planted in a single season. However, it would seem very unlikely based on the following factors:

1. The upland Trannon catchment is only 55 % forested (the Coalburn catchment is 100 % forest)
2. Planting of Dolgau forest occurred progressively over a 30 year period (the entire Coalburn catchment was prepared and planted in a single season)
3. Felling occurs in discrete forest 'parcels' meaning that at any one time only a small proportion of the forest is without canopy cover

4. The buffering effect of non-forested tributary inputs upstream of the lowland reaches would act to reduce the impact of Dolgau forest hydrologic runoff

Rather, it seems likely that upland catchment forestry will have altered lowland peak discharges only slightly at Trefeglwys, and almost certainly not enough to produce the instability observed in the period 1963-1976 or the increased magnitude and frequency of competent flows since 1989.

8.4 Stable lowland reaches – possible reasons for stability

Not all lowland reaches of the Trannon have experienced the high rates of channel change experienced in the Trefeglwys study reaches. In particular, the Ddranen Ddu study site serves as an example of a lowland reach which has experienced no significant lateral movement or channel widening in the contemporary study period and where visual analysis of historical aerial photographs suggests that the reach has also been stable throughout the medium-term.

The reach, which is 1.4 km downstream of Bodiach Hall, is characterised by low sinuosity (1.0), low bed gradient (0.003 – 0.006) virtually no apparent bed load storage and cohesive silt / clay banks containing no gravel. Where eroded, the banks at Ddranen Ddu have undergone rotational slumping, however, the channel differenced DTMs demonstrate that the bank erosion rate is extremely low in the reach.

In light of the above discussion there are several explanations for the stability of the Ddranen Ddu reach, which are by no means mutually exclusive:

1. That upstream inputs to the reach, resulting from upland afforestation and local bed load inputs from bank erosion in the Trefeglwys reaches, have not been transported downstream as far as the Ddranen Ddu reach. Consequently, it is transporting far lower volumes of bed load than the channel at Trefeglwys .

2. That the silt / clay bank material at Ddranen Ddu means that local additions to the coarse sediment budget of the reach are extremely small, again meaning that the volume of bed load having to undergo transport through the reach is low.
3. That the cohesive nature and low sinuosity of the Ddranen Ddu banks combine to mean that even high flows cause little bank erosion.

It has been outside of the remit of this study to quantitatively examine the relative merits of the above statements, however the bed load transport rates from the Bodiach Hall reach would suggest that only a very small proportion of bed load inputs to the Trefeglwys reaches are transported out of the reach, with the majority quickly entering local storage. In response, the opposing bank erosion ensures channel volume remains relatively stable. At the same time, local coarse sediment inputs to the Ddranen Ddu reach are almost non existent because there is no gravel present within the banks and the banks have a much greater resistance to erosion from high flows than the composite banks at Trefeglwys. The much greater resistance of silt / clay banks is supported by the findings of Winterbottom and Gilvear (2000) who found that in the River Tummel, Scotland, composite banks eroded at more that ten times the rate of silt banks.

In summary it would seem that all of the above three factors have combined to promote stability at Ddranen Ddu, although determining the relative significance of each factor is highlighted as an area for further work.

8.5 A conceptual model for the influence of upland catchment and local factors on the lowland channel changes of the Afon Trannon

The preceding discussion emphasises that the medium-term instability of the lowland Trannon can be linked to:

1. Elevated upstream bed load inputs resulting from Dolgau forest
2. Local bed load inputs from composite bank erosion

3. The downstream change in valley character

And that:

4. More recently increased magnitude and frequency of competent flows may have elevated channel change rates further.

Figure 8.12 presents a summary conceptual model, based on the findings of this project and their subsequent discussion, which explains the medium-term destabilisation of the lowland Trannon, and in which the influencing factors combine to produce such a feedback mechanism. It also implies the importance of the downstream distribution of bank material on reach stability and predicts that in locations where composite bank material exists, providing large volumes of local bed load inputs, the reach will be unstable. Conversely, where the bank material is cohesive and contains little or no gravel upstream bed load inputs will be transported quickly through the reach with little resultant instability.

Such a pattern of stable and unstable reaches is similar to that described on the Bella Coola River, Canada, by Church (1983) where unstable 'sedimentation zones' are separated by stable 'transfer reaches'. In the Bella Coola River the downstream locations of sedimentation zones are controlled by local bed load inputs from tributaries which form alluvial fans which act to reduce the local bed load transport rate. In the Trannon the situation is slightly different but somewhat analogous. Here, the downstream pattern of stability is controlled by the bank composition with 'sedimentation zones' occurring in reaches with composite banks and 'transfer reaches' occurring where cohesive bank material acts to promote stability. Local, lateral inputs from bank erosion are, in many ways, similar to Church's local tributary inputs, except that they cause instability not by their ability to restrict local bed load transport rates, but rather by their redeposition onto downstream bar forms where they cause further bank erosion and local bed load inputs.

The conceptual model, in which instability is triggered as a result of local aggradation following forestry, provides one explanation of lowland channel destabilisation in the Afon Trannon. However, it should be noted that it does not fully explain the pattern of aggradation and locations of instability seen in chapters 6 and 7. Indeed, it prompts the following question:

'Why is the aggradation mainly restricted to the meander bends?'

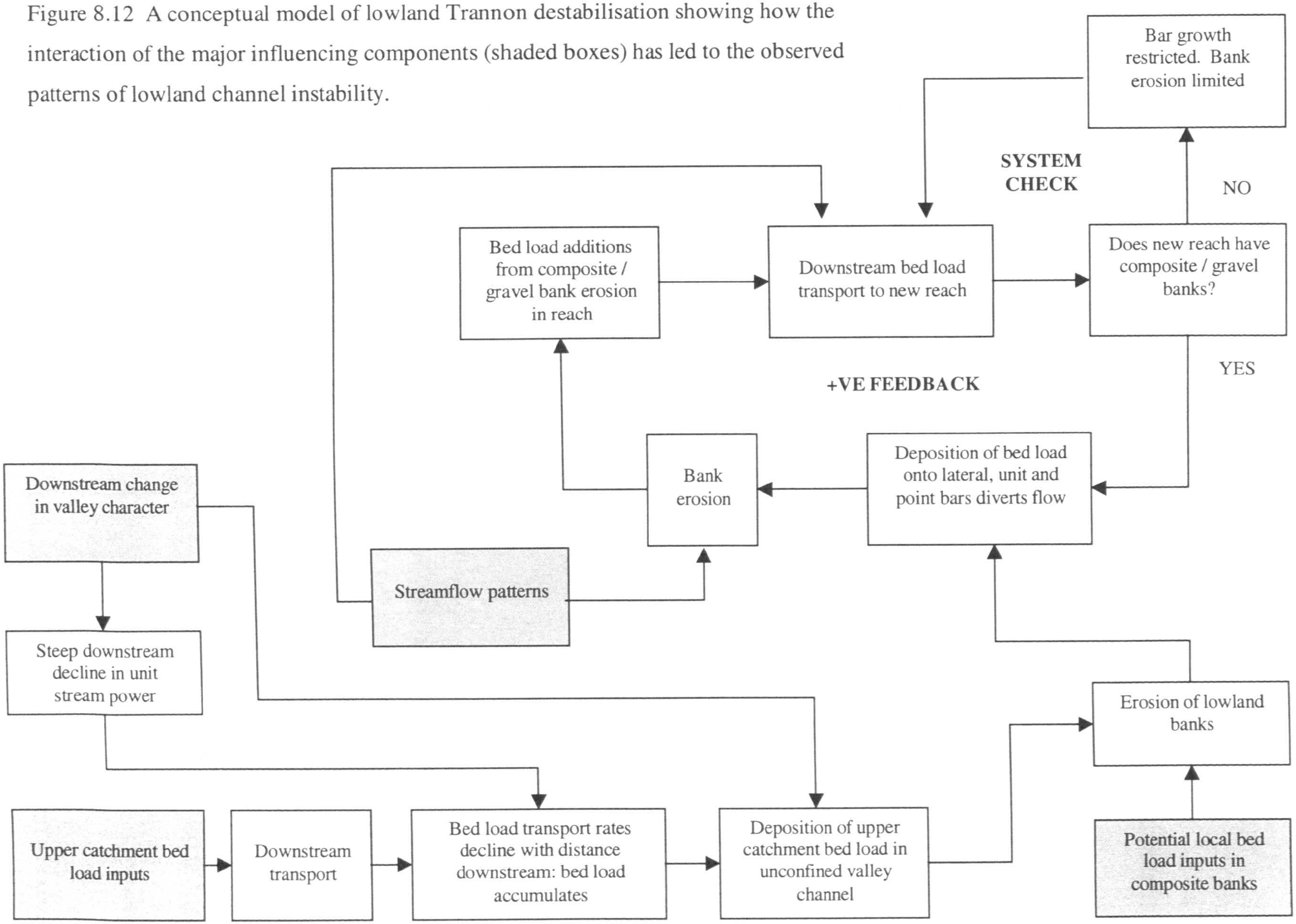
A further possible explanation is linked to the meander radius of curvature, channel width and rates of bank erosion (c.f. Hickin and Nanson, 1975; Hickin and Nanson, 1984). Markham and Thorne (1992) divided the meander cross section into three regions:

1. Mid-channel region: 90 percent of the flow passes through this region and the classic helicoidal motion is well established
2. Outer bank region: The cell of opposite circulation develops in this region. Importantly, the strength of this cell increases with discharge, the steepness of the bank and the acuteness of the bend.
3. Inner bank region: where it is argued that shoaling over the point bar induces a net outward flow forcing the core of maximum velocity towards the outer bank.

The interaction of the outer bank flow and the channel banks themselves produces bank erosion and it is clear from Markham and Thorne (1992) that where meander steepening occurs, so the rate of bank retreat increases. This said, meander bank erosion could therefore be triggered in the lowland Trannon wherever there happens to be a meander bend that has grown fairly sharp, causing local aggradation downstream.

Determining to what extent such a mechanism has occurred on the Trannon represents an interesting area for further research and would allow the validity of the conceptual model presented in figure 8.12 to be better affirmed.

Figure 8.12 A conceptual model of lowland Trannon destabilisation showing how the interaction of the major influencing components (shaded boxes) has led to the observed patterns of lowland channel instability.



8.6 Broader implications for river management

8.6.1 Implications for the management of UK rivers with upland catchment forestry

The recent increases in UK upland catchment afforestation have been geographically wide ranging. Although the main concentration has been in Scotland (up to 35,000 ha yr^{-1}), significant afforestation has occurred in the northern counties of England (up to 12,000 ha yr^{-1}) and Wales (6000 ha yr^{-1}). Consequently, the medium-term impacts of upland catchment afforestation are far from being local issues, instead they have the potential to affect many rivers draining upland areas located throughout the UK. Consequently, the findings of this study provide a useful resource for the managers of the numerous rivers draining forested upland catchments.

The current estimated bed load yield from the upper Trannon catchment and the % forest cover is compared to that from other catchments containing mature forest in the UK in table 8.11.

It is clear from the table that:

1. The bed load yield from forested upland catchments is extremely variable, even when the catchments are in a similar geographic location and have the same underlying geology (i.e. Tanllwyth and Hore which are adjoining catchments).
2. The bed load yield of the upper Trannon catchment is low compared to other Welsh catchments (in the main part due to its lower % cover and low slope angle), and also lower than that of Kirkton Glen, despite Kirkton Glen having a lower percentage forest cover than the Trannon.

Table 8.11 Comparison of bed load yields of the Hore, Tanllwyth Kirkton Glen and Trannon catchments, UK.

Catchment	**% of catchment forested**	**Bed load yield ($t\ km^{2}\ yr^{-1}$)**
Hore, mid-Wales (Leeks, 1992)	100 %	11.8
Tanllwyth, mid-Wales (Moore and Newson, 1986)	100 %	41.29
Kirkton, Scotland (Johnson, 1993)	40 %	6.85
Trannon, mid-Wales	56 %	5.24

Much of the above variability may be explained by the differing physiography of the various catchments. Kirkton Glen, for example, is particularly steep (mean slope angle is 22° compared to 14° in the Tanllwyth catchment and just 7° in the forested regions of upper Trannon catchment). This variability in bed load yield from forested catchments is an essential consideration if one is to attempt to assess the likely impact of upland afforestation on a lowland channel. Essentially it means that there is no 'golden rule' which allows an accurate estimate of bed load yield to be generated for a catchment, instead each upland catchment must be assessed separately, in terms of its physiography, geology and forest cover, in order to gain a meaningful estimate of the bed load yield one might expect to lowland channels. However, the following rules are generally applicable:

- The steeper the catchments the greater the potential for high bed load yields
- The greater the percentage forest cover the greater the bed load yield
- The larger the catchment the lower the bed load yield per unit area

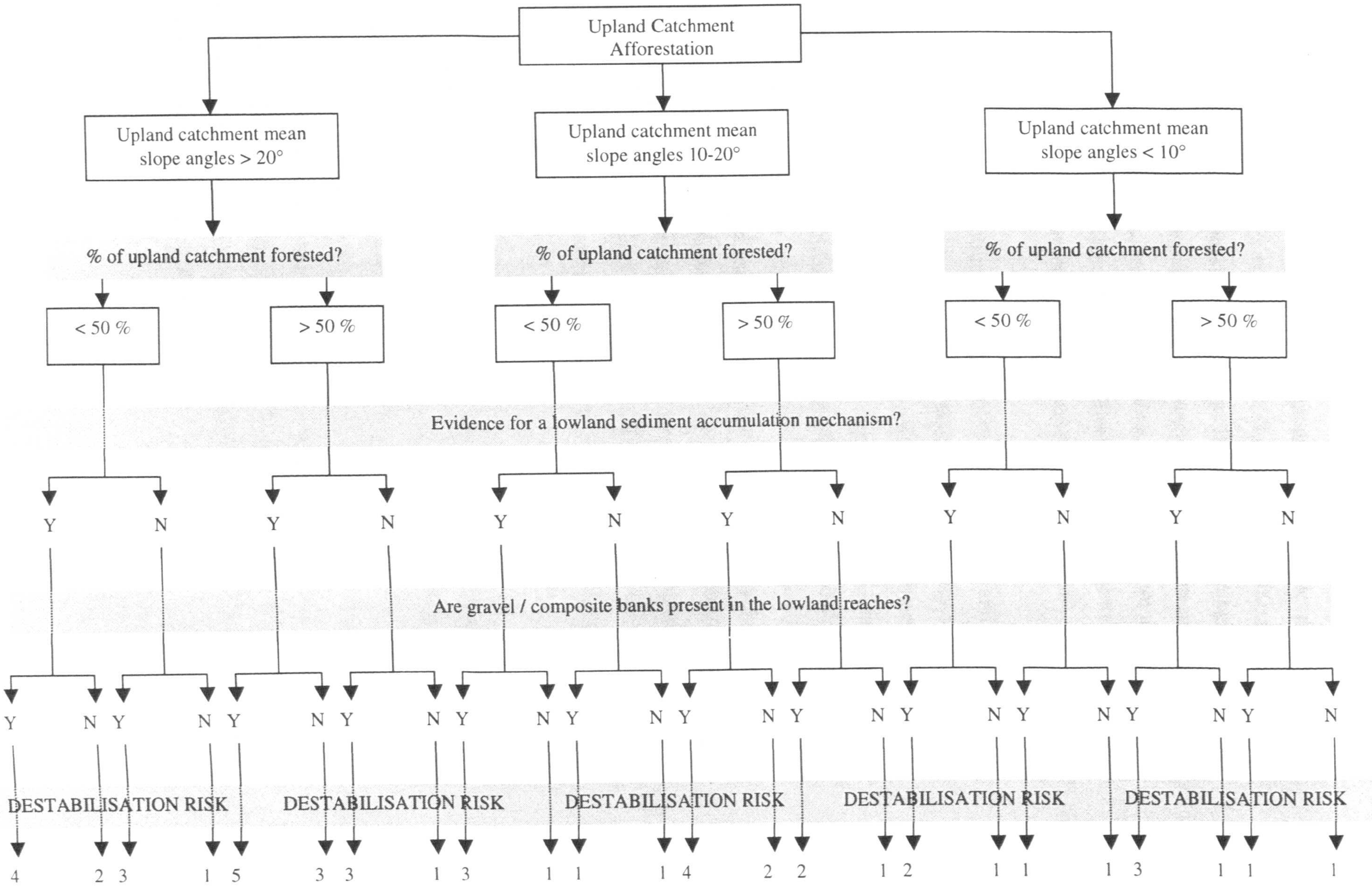
Therefore, the lowland channels (with upland catchment forestry) at the most risk of destabilisation are those which drain many small, steep, upland catchments which have a high percentage forest cover. In mid-Wales the majority of upland catchment afforestation has occurred on the relatively flat Cambrian Upland plateau. The upland catchments of the River Severn and River Wye, for example, seldom have slope angles greater than 15° (Kirby et al., 1991). Consequently, one might expect the forestry of mid-Wales catchments to generate bed load yields of a magnitude incapable of producing lowland channel destabilisation as a direct causal link and in the case of the Afon Trannon this is true. Instead, this study emphasizes the importance of lowland channel feedback mechanisms which may be triggered by relatively small changes in the upland catchment bed load yields, in this case local elevations in composite bank erosion. Consequently, channels near their threshold of stability are prime candidates for lowland channel destabilisation following upland catchment afforestation.

By means of a summary, figure 8.13 represents a simple tool by which the potential risk of UK lowland channel destabilisation following upland catchment afforestation may be estimated. It emphasizes the importance of catchment physiography and forest cover, the need for lowland sediment accumulation mechanisms and the important contributions of lowland bank material in the production of a risk value. In the context of the flow diagram the Afon Trannon represents a medium risk river.

8.6.2 Promotion of a holistic, catchment-scale approach to river management

Traditionally there has been a friendly rivalry between river engineers and geomorphologists which is due, in part at least, to the different approach to the study and management of river systems taken by each party (Newson, 1995). River engineers commonly attempt to solve point problems via local management strategies, whereas geomorphologists tend to approach the study of fluvial systems from the catchment-scale. Recently the approaches of engineers and geomorphologists have become more integrated (Sear et al., 1995). Much of this

Figure 8.13 Risk of lowland destabilisation as a result of upland catchment afforestation. Low risk = 1, medium risk = 3, high risk = 5.



integration can be linked to the increasing interest in 'sustainable' management in which river management is expected 'to last' (Leuven et al. (2000) rather than be a short-term solution to a longer-term problem. Engineers increasingly recognize the importance of information gained from large-scale, medium-term and longer-term geomorphological approaches to understanding the fluvial system which are essential facets in the design of river management strategies able to last. Put simply, a holistic approach to river management is increasingly 'in vogue'.

This research supports a holistic approach to river management and the need to approach catchment-scale fluvial problems from an ergodic perspective. In particular it highlights the way in which the action of lowland channels may be a combined response to both local factors (i.e. bank material) and those upstream (i.e. valley form and upland land use change), each requiring consideration. It also highlights the importance of long-term controls (i.e. the valley form) medium-term controls (i.e. upstream bed load inputs from land use change) and short-term controls (i.e. variability in the magnitude and frequency of competent flows). As such it strongly supports the views of Lane and Richards (1997) who argue that the recent reduction in the time and space scales of fluvial research has meant that continued advances in the understanding of fluvial system evolution is in danger of being restricted.

Also highlighted by this study is the way in which abrupt changes in the above controls may cause an abrupt change in channel form and processes, particularly where change occurs in more than one control. Consequently, it is the author's belief that the early search for, and field identification of, such abrupt changes would greatly benefit the appropriate design of sustainable lowland river management schemes. Consequently the following points, identified in this study, are offered as features which may be quickly identified in the field, and which may be extremely influential in terms of local channel stability:

- Long-term, catchment-scale factors

  - Abrupt decrease in valley gradient
    May cause a sharp decline in unit stream power and hence a decrease in bed load transport rates. Expect local, within channel storage of sediment, reduction of channel capacity, possible channel instability and increased flood hazard.

  - Abrupt large increase in valley / floodplain width
    Indicates a change from bed rock to alluvial floodplain controls and hence the ability of the channel to adjust laterally. Expect lateral storage of bed load, bank erosion and local sediment inputs. An increased supply of sediment from upstream has potential to increase lateral change rates.

- Long-term, reach-scale factors

  - Abrupt change in bank material from sand / silt / clay to composite and gravel banks. Indicates an increased potential for lateral inputs and a reach in which stability may be marginal. Expect elevated rates of channel change. Increased upstream sediment supplies and increased flow magnitude and frequency may cause significant instability.

- Medium-term catchment-scale factors

  - Medium-term increases in upstream sediment supply (i.e. mining, upland catchment forestry). If sediment yielded is in sufficient quantity channel may experience a direct response. Expect increased bed elevation, increases in channel width, decreases in channel depth, increased instability. Even when relatively small quantities of sediment are yielded, may combine with longer-term controls to produce instability via indirect feedback responses.

The potential for such responses is more difficult to identify. Problems are most likely to occur where one or more controlling factors combine – i.e. where storage is promoted by a change in valley character and where the existence of composite bank material has the potential to provide large local sediment inputs.

- Increase in flood magnitude and frequency. Expect increased channel instability especially in reaches where composite bank material is present.

9. CONCLUSIONS, APPRAISAL OF TECHNIQUES AND FURTHER WORK

9.1 Summary of main findings

The overriding aim of this project was to determine the extent to which upland catchment afforestation of the Afon Trannon can be held responsible for its medium-term and contemporary lowland channel instability, and whether lowland channel changes can be thought of as a direct or indirect response to upland land use change. On the basis of this work, the following general conclusions can be reached:

- Lowland channel destabilisation of the Trannon is not a direct response to elevated upland catchment bed load yields caused by upland afforestation. The estimated medium-term quantity of bed load yielded from the upland catchment entering the lowland sediment budget is small compared to the estimated bed load yield derived from lowland bank erosion. Both in the medium-term and contemporary sense, lowland bank erosion, rather than upstream supply, forms the dominant process of bed load yield to the lowland channel reaches.

- Although lowland channel destabilisation of the Trannon is not a direct response to elevated upland catchment bed load yields caused by upland catchment afforestation, there is circumstantial evidence to indicate that upland catchment forestry may be an indirect factor. In order to implicate upland catchment bed load, a triggering mechanism is required whereby local lowland bed aggradation causes lowland bank erosion capable of initiating a positive feedback such as that presented in the conceptual model (figure 8.12). It is thought that the coincidence of a change in the valley character and bank material above Trefeglwys may promote such a mechanism, however it is also recognised that other lowland channel change mechanisms (i.e. elevated bank erosion following meander tightening) may also be of importance.

- The percentage of the river catchment which is afforested is of clear importance in respect of the resulting bed load yields and hence the potential for destabilisation. Therefore, the percentage of the catchment area which is deemed 'upland' is also of importance. In the Afon Trannon catchment, where the upland catchment is extensive (15 % of the total catchment area) 8.3 % of the total catchment area is forested, yet a direct link between upland catchment afforestation and lowland destabilisation is negated. It is the author's opinion that only in catchments with exceptionally large upland areas, in which the total upland area was afforested, would the potential for lowland destabilisation to occur as a direct consequence of bed aggradation in response to elevated upland bed load yields exist.

- Overall catchment physiography and the length of the transfer reach are also important factors in determining the quantity and timing of bed load transferred to the lowland channel.

- The nature of lowland bank material is a dominant control of lowland bank erosion rates, and hence the potential for local bed load inputs from lowland banks. The risk of indirect lowland instability, via feedback mechanisms triggered as a result of upland catchment afforestation, is enhanced considerably where lowland banks are composite in nature, eroded easily and are composed of a high proportion of material of bed load size. It would be unlikely for the stability of lowland channels with fine, cohesive banks to be seriously affected by elevated upland catchment bed load yields resulting from upland catchment forestry.

- The influence of competent flow magnitude and frequency on lowland channel change rates is well demonstrated with an apparently strong relationship existing between the maximum stage reached in each survey period and the total bank retreat. As a result, the recently high rates of lowland channel instability are undoubtedly strongly influenced by the increase in flood magnitude and frequency in the period 1988-2000. In this case it would appear that the cause of the increased magnitude and frequency of competent discharge is a result of climatic change rather than an impact of forestry, although further investigation of climatic data is required to confirm this view.

9.2 Appraisal of the approach taken

The nature of this project has necessarily required a medium-term, catchment-scale view of the Trannon to be taken. This has produced a study in which the dominant land use change and resulting macro-scale fluvial processes acting within an area of 72 km^2 have been investigated over a 50 year period. Such catchment-scale approaches to fluvial geomorphology are becoming less and less common in the literature with a trend towards small-scale, reductionist studies having developed. In such studies elements of the fluvial system are investigated in short study reaches and commonly over short-term study periods. Commonly, the number of parameters which require investigation to allow the identification and quantification of the processes about which information is required is constrained by the often small-scale nature of the processes themselves meaning that the required data can fairly easily be acquired via field study. In this study, however, the opposite is true. Investigation of the dominant catchment and channel processes acting at a large temporal and spatial scale, and able to be quantified from limited historical records, have been investigated with smaller-scale processes (i.e. those acting to produce micro and meso-scale changes in channel form) forming a lesser element of the study. Commonly, high quality historical data have been scare and data have had to be generated using empirical formulae or modelling approaches in which the data accuracy is not fully tested.

Such large-scale research projects present many problems, not least the need to correctly identify the important parameters for study amongst the almost endless number within large catchments. But of fundamental constraint is the logistics of researching large systems over extensive temporal periods. Logistical issues come to the fore when the research is conducted by a team of small size, or a single individual as in the case of this research. Such a small number of researchers necessarily limits the volume and type of data which can be acquired. Consequently, it is necessary to appraise products of this research in terms of the extent to which they have met the project objectives. The appraisal is presented in Table 9.1.

The accuracy of the estimated upland catchment bed load yield can be debated based on whether it is agreed that bed load yield is a catchment physiography or channel property. The rainfall – runoff modeling appears to have worked well providing a data set which appears to show recent climatic change. Aerial photogrammetry was limited in older images, however the accuracy achieved was of a magnitude which did not prevent useful analysis. Finally, the use of a GIS in the production of channel DTMs represents a significant advance in fluvial geomorphology and demonstrates that traditional section data may be well analysed by the use of DTMs.

9.3 Further work

This study represents a first step towards determining the medium-term impacts of UK upland catchment forestry on the processes acting in the lowland channels and their stability. It should by no means be considered complete, rather, in respect of the currently extremely limited range of UK-based literature examining the issue, it forms an important starting point from which further studies may continue. It has identified many of the factors influencing the lowland stability of the study site, however it is limited by its case study approach. Further medium-term studies of a wide range of UK catchments with upland afforestation are needed to generate a more extensive data set from which more general conclusions may be drawn.

The study has raised a large number of questions which have not been fully answered by the adopted approach and which would benefit from further work.

Table 9.1 Appraisal of the techniques used to satisfy objective 1.

Objective	Appraisal of the product used to satisfy the objective
The generation of a medium-term upland catchment bed load yield for the Afon Trannon	Whilst the sediment delivery ratio technique used did allow the simple production of a bed load yield estimate for the upper Trannon it is limited by the extent to which bed load yield can be thought of as a physiographic property or a channel property. The approach adopted assumes bed load to be a product of catchment physiography, however in reality it is also strongly influenced by channel properties (i.e. discharge and the bed load transport rate). The error inherent in the approach taken is extremely difficult to quantify unless an independent record of bed load yield can be established for the upland Trannon catchment (i.e. using bed load trapping as in the Tanllwyth).
The generation of a medium-term hydrologic record for the Afon Trannon	The data-based mechanistic model used represents possibly the only way to estimate historical discharge in the lowland Trannon with reasonable accuracy due to the extremely limited data available for the catchment. The model validation to measured data suggests that the technique has been successful with nearly 80 % of the contemporary flow data explained by the model. The major weaknesses of the flow data are: 1. Without a rating curve the stage record can not be converted to discharge for use in bed load transport formulae etc. 2. The length of the model output is restricted by the length of the rainfall data used as the model input
The quantification of historical rates of change of the Afon Trannon	The main restriction of the methodology employed has been the quality of the aerial photography used. As a general rule, the earlier the photography the poorer the quality of the photography. Where ortho-correction procedures have been employed the resulting data is of a quality capable of allowing the analysis of two dimensional channel changes, however it was incapable of allowing investigation of channel change in the Z direction. Where geo-correction was employed the error was slightly greater than that of ortho-corrected imagery but still of an accuracy capable of identifying channel change. Despite rectification, the photogrammetric techniques still contained significant error and it is unlikely that such an approach would produce useful results if used on river channels of much greater stability than the Trannon.

Objective	Appraisal of the product used to satisfy the objective
A contemporary channel change record allowing examination of the processes of channel change	Much of the contemporary assessment of the lowland channel has revolved around cross profile data. Whilst this type of data source is well established in the literature, the use of a GIS to produce an interpolated surface from which differenced DTMs can be produced is relatively new. Arguably, the best interpolated surface is generated when the survey point resolution is similar in both a cross stream and downstream direction (e.g. Brassington et al., 2000). However, the ability of the DTMs in this project to represent meso-scale bed form well at Bodiach Hall suggests that the approach can be applied to more traditional cross profile data with good results, especially if the downstream resolution of the sections is high. This is an important finding and suggests that the analysis of older cross profile data sets may benefit from reappraisal using surface interpolation approaches, particularly where the interest is in meandering channels.

a) Further data collection

1. Independent determination of the medium-term nature of bed load transport through the middle transfer reaches of the Trannon is required to determine the lag time between upland catchment afforestation and lowland channel response. The generation of such data represents a complex problem and would require historical bed load transport rates to be determined at a number of locations through the transfer reach using bed load transport data acquired in the field (i.e. Helly-Smith type sampling) or that generated using empirical formulae.

2. Further analysis of lowland channel planform development, in particular historical assessment of meander radius of curvature to determine whether the initial local bed aggradation required to trigger the positive feedback loop presented in the conceptual model may have been produced by the natural sharpening of meander bends. This could be simply achieved via analysis of rectified aerial photography and represents a relatively simple addition to this work.

b) Methodological issues

3. The accuracy with which bed load yield estimates may be generated using empirical sediment delivery ratio formulae has not been addressed despite its previous use in the assessment of bed load delivery (Newson, 1980). One way of approaching the problem might be to collate all available bed load yield data from monitored catchments for which catchment physiography data exists and compare the measured data to predictions for the catchments based on empirical formulae.

4. The study has highlighted the value of surface interpolation as a means of generating channel DTMs form which channel morphologic change may be determined and the significant advantages of adopting such an approach over a

traditional cross profile analysis approach However, the project has also highlighted the difficulty in determining the error inherent in such interpolations. Commonly, the issue has been largely ignored, with little attention given to the limitations of the acquired surfaces. However, with the use of such techniques set to become common place in fluvial geomorphology there is a pressing need to address the issues of interpolation error in river channels and, where cross profile data is used, the downstream and cross stream resolution required to produce a surface of acceptable error.

5. This study has utilised a TIN approach to surface interpolation, however there are several ways by which vector data may be interpolated including kriging, nearest neighbour and inverse distance methods. It would be a useful extension to this project to undertake surface interpolation via other methods with a view to determining the best way to interpolate channel survey data.

c) Broader findings

6. Of particular interest is the recent increase in the magnitude and frequency of competent flows highlighted by the ranifall – runoff modelling which appear to be climatically induced. Currently, short-term climatic changes and their impact of flood discharges are of particular public interest. Therefore, the investigation of recent climatic change in mid-Wales, particularly the temporal pattern of rainfall, would form an extremely relevant extension to this work.

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